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CONTENTS

PAGE

WILSON, C. B. The Upper Middle Hecla Hoek Rocks of Ny Friesland, Spitsbergen 89

CARTER, CAROL S., and RICHARD J. CHORLEY. Early Slope Development in an Expanding Stream System 117

MCCANN, S. B. Some Supposed "Raised Beach" Deposits at Corran, Loch Linnhe, and Loch Etive 131

PHILIP, G. M. Lower Devonian Crinoids from Toongabbie, Victoria, Australia 143

SELWOOD, E. B. The Upper Devonian and Lower Carboniferous Stratigraphy of Boscastle and Tintagel, Cornwall 161

CORRESPONDENCE
pp. 168–173

REVIEW
p. 174

PUBLICATIONS RECEIVED
p. 174

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GEOLOGICAL MAGAZINE

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The Upper Middle Hecla Hoek Rocks of Ny Friesland, Spitsbergen

By C. B. WILSON *

(PLATES IV–VII)

ABSTRACT

This paper gives a definitive description of the lithology and thicknesses of the Upper Middle Hecla Hoek rocks of Ny Friesland, Spitsbergen, arising out of Cambridge expeditions since 1951. The succession is the late Pre-Cambrian part of the Hecla Hoek Geosyncline, lying below the Tillite horizons and characterized essentially by limestones and dolomites.

I. INTRODUCTION

(*a*) *Historical*

THIS paper follows one describing the Lower Middle Hecla Hoek (Wilson 1958), which contains place-name maps to which reference should be made. The Hecla Hoek succession of which the present paper describes a part was first outlined in 1956 (Harland and Wilson).

The Upper Middle Hecla Hoek comprises the great calcareous series broadly equivalent to the Upper Murchison Bay Formation of Nordaustlandet and the Upper Eleonore Bay Formation of east Greenland, all late Pre-Cambrian in age. During several Cambridge summer expeditions the Ny Friesland succession was investigated, resulting in good correlations throughout the peninsula except, perhaps, in the extreme north where further work is required. Following W. B. Harland's observations in 1951 (Harland and Hollin 1953) the author accompanied the British Spitsbergen Expedition 1952 (Bayly 1953), working between Oslobreen and Oxfordbreen, and used material collected by M. B. Bayly, P. B. H. Bailey and K. C. Stileman, but the following account derives mainly from his own work during the summers of 1953 (Harland 1954), 1955 (Harland 1957) and 1956 (Wilson 1957).

Stratigraphical and structural studies were closely related. A preliminary account of this structure was given by Harland (1959).

(*b*) *Succession*

The rocks are stratigraphically of interest on account of their wide variety of lithology and colour and also the complex sequences found at certain levels. It is possible to make reliable correlations on a

* Just before his death in 1959 Mr. Wilson was working on the MS of this his most important contribution. It has been cut and edited but without loss of the many detailed observations from several arduous journeys in a remote region. W. B. H.

lithological basis, both in the case of thin bands and sequences of contrasted rock types and also on the broader scale of rock formations. The general succession in southern Ny Friesland is as follows:—

The Akademikerbreen Series	(*d*) Backlundtoppen Dolomites (Three Members, unnamed)
	(*c*) Backlundtoppen Oolites (Three Members, unnamed)
	(*b*) Draken Conglomerates (undivided)
	(*a*) Svanbergfjellet Beds: 4. Upper Svanbergfjellet Limestones. 3. Svanbergfjellet Collenia Beds. 2. Lower Svanbergfjellet Limestones. 1. Svanbergfjellet Dolomites.
The Grusdievbreen Series	(*b*) Upper Grusdievbreen Limestones (Six Members unnamed)
	(*a*) Lower Grusdievbreen Limestones (Four Members unnamed)

The general distribution of these formations and the marker horizons used in correlation are shown in Table 1. Most formations are recognized by their combined lithology, thickness, and succession, but the following transitions and marker horizons have proved especially useful.

TABLE 1.—THE DISTRIBUTION AND THICKNESS OF UPPER MIDDLE HECLA HOEK FORMATIONS IN NY FRIESLAND.

	Formation.	Akademikerbreen (A,P,±⅛)	Grusdievbreen (P,M,±⅛)	Polarisbreen (P,M,±¼)	Raudberget (P,M,±¼)	Glintdalen (F,M,±¼)	N.W. Lomfjorden (F,M,±½)	Kluftdalen (F,±½)
AKADEMIKERBREEN SERIES	(d) B.D.	300	300	700	250	300	300	150
	(c) B.O.	350	350		350	350	375	210
	(b) D.C.	300		180	170	215	160	25
	(a) S.B. 4	115			40			3
	(a) S.B. 3	230				150	220	52
	(a) S.B. 2	130				40	55	15
	(a) S.B. 1	150±		1250		160	75	30
GRUSDIEVBREEN S.	(b) U.G.L.	465			(Breccia bed).	400	400	?400
	(a) L.G.L.	400 (Rogatsjevryggen)			400	400	350	2–500

The symbols in brackets indicate the method of estimation (A, aneroid; P, photos; M, map; F, pacing), and the fractions the maximum probable errors due to all causes (mechanical deformation, irregularity or uncertainty of dip, and errors of measurement). The thick vertical lines indicate the continuity of exposure (diagrammatic), and the arrows show a continuation into neighbouring series. Formational names are abbreviated.

(i) The lower boundary of the Grusdievbreen Series marks the abrupt appearance of purely calcareous facies with a pseudo-conglomerate at the base above the very thick and generally quartzose rocks of the Lower Middle Hecla Hoek, a level which may prove to be of wide application in Spitsbergen. The boundary between the two

series cannot yet be specified to within less than 50 m., owing to an apparent alternation of facies, but it accords as well as possible with that between the Hunnberg and Ryssø Series of Nordaustlandet since it marks the first appearance of thick dolomite beds, including those of Ryssø Dolomite type.

(ii) The transition from the Lower to the Upper Grusdievbreen Limestones, with the red flags and conglomerates. The Upper Grusdievbreen Limestones alone appear to have unique characteristics.

(iii) The transition from the Svanbergfjellet Dolomites to the Lower Svanbergfjellet Limestones, with the " Bolster Beds ".

(iv) The Svanbergfjellet *Collenia* Beds.

(v) The transition from the Backlundtoppen Oolites to the Backlundtoppen Dolomites with the " Coloured Oolite Horizon ".

(vi) The top of the Akademikerbreen Series with the " Dartboard Dolomite " is everywhere clearly defined at the base of the thick dark shaly beds of the Polarisbreen Series * (included by Kulling, 1934, within his Ryssø Series).

The formations are described in ascending order, each from south to north, with petrographical details where these are especially relevant. Other petrographical observations and comments on the various sedimentary structures are generalized in Section IV, being based on thin sections representing only about a third of the available specimens.

Cryptozoon or Stromatolitic Structures are here termed *Collenia* (see Section IV(*d*) and Text-fig. 2). *Oolites* and *pisolites* are respectively less than and more than 5 mm. in diameter. *Oolitoids* are described in Section IV(*c*).

II. The Grusdievbreen Series

The Series is composed mainly of limestones which are separated throughout Ny Friesland into an upper pale and a lower dark formation. In the south there is a thin, dark red horizon, separating the pale from the dark facies, which is taken as the base of the Upper Limestones. The series has only been fully investigated in the south—the upper part in Akademikerbreen, and the lower part in the Grusdievbreen area on either limb of the " Oxfordbreen syncline ". These sections overlap, and between them they cover the complete series.

(*a*) *The Lower Grusdievbreen Limestones*

Complete and undisturbed exposures are rare, but the following general subdivision is based on observations of a large number of

* The author was collaborating with W. B. Harland on a fuller description of these rocks.

partial exposures throughout central and southern Ny Friesland which indicate a uniform stratigraphy:—

4. 3–15 m. Dolomite, weathering pale.
3. ca 100 m. Limestone, black.
2. ca 275 m. Dolomitic limestone shales and flags, nodular, dark.
1. 25–30 m. Dolomite, weathering pale.

Member 1 was found in Transparentbreen, Oxfordbreen, Skålbreen (south of Glintbreen), and possibly Faksevågen, and has nowhere been proved absent. On Rogatsjevryggen in Transparentbreen (where the whole formation is exposed in a sharp arête) it consists of a well defined band, 30 m. thick, of orange-weathering, dark grey dolomite flake-conglomerate, a resistant rock which forms a prominent subsidiary peak in the ridge (much as do similar bands in the Tillite Formation). In Oxfordbreen the division consists at least in part of cream-yellow-weathering, dark grey dolomite-siltstone flags with very smooth bedding planes. In Skålbreen, 15 m. of orange-brown-weathering dolomite pebble-conglomerate are overlain by 10 m. of dolomite-siltstone of the same lithology as in Oxfordbreen, each band sharply defined above and beneath. At the base of the Faksevågen cliff Kulling (1934) noticed but did not examine 20 m. of pale-weathering rock which, falling at exactly the same horizon, most probably represents the same band.

Member 2 is most characteristic of the lithology of the formation and is a unique marker horizon. The basal beds are excellently exposed at the eastern end of the Faksevågen cliff (Wilson 1958, Plate VII). The beds are of a pale appearance due to a higher degree of dolomitization than is usual in this division. The rocks consist of a very striking pseudo-conglomerate, the like of which was not seen at any other level in the Hecla Hoek. Thin layers of buff-weathering, dark grey dolomite, 2–15 cm. thick, alternate with similar layers in which the dolomitization has left pancake-shaped areas of the original limestone-siltstone unaffected (Plate IV, fig. 1). Individual plates are commonly 2–10 cm. thick, and are of such a shape as to suggest desiccation without actual disruption; thus they tend to have depressed centres and down-warped margins. The boundary between dolomite and limestone is usually transitional, often by very fine lamination between the two, a feature which is also common in the normal bedded rock. The plates tend to have an outer margin of pure limestone with a central area which is partly dolomitized. Both dolomite and limestone show well developed false bedding.

The greater part of the division is of a similar lithology but of less extreme development, so that the limestone layers are not split up into plates but form continuous undulating or nodular bands, alternating with dolomitic layers. There is also much intercalated shaly limestone,

and in Grusdievbreen there is evidence of considerable replacement of some bands by black chert, giving a typical nodular rock on weathering. Pyrite is abundant on some bedding planes.

Member 3 is more uniform and thickly bedded than Member 2, and consists of black-weathering, black limestone-siltstones with false bedding and some layers of dark flake conglomerate (see Section IV(*b*)).

Member 4 is known from Grusdievbreen, Fulmarberget and Akademikerbreen (Golitsynfjellet), and these correlate roughly as follows:—

Grusdievbreen and Oxfordbreen.	Golitsynfjellet.
e. Thick limestones, banded cream and medium grey.	*e.* Thick limestones, banded cream and medium grey.
d. 8 m. Limestone, massive, white.	*d.* 10 m. Limestone, massive, pink-white, with *Collenia* and flake-conglomerate.
c. 5 m. Calcareous flags, purple-brown, with intra-formational conglomerate.	*c.* 6 m. Calcareous flags, purple-brown, with intra-formational conglomerates.
b. 1½ m. Dolomite-mudstone flags, grey, weathering pale, laminated.	*b.* 3 m. Dolomite pebble-conglomerate, grey, weathering cream-orange.
a. Limestone flake-conglomerate, partly dolomitized, as thin (¼ m.) layers within the top few metres of the black limestone of Member 3 type.	*a.* Thin conglomerate layers, identical with those of Oxfordbreen, within the top few metres of the black limestone of Member 3 type.

Member 4 is represented by *a.* and *b.* above. The flake conglomerates of *a.* are of interest in that they differ in appearance from those found in other formations. Thus in Oxfordbreen and Akademikerbreen they are of two sorts, one with pale brown-weathering dolomitized flakes in a pure white calcite matrix (Plate VII, fig. 5), and the other with grey flakes in a similar matrix. The lithology of *b.* is similar to that of Member 1, and further observations may establish it in other areas.

Other good exposures of the formation as a whole occur around Dracoisen, Raudbergbreen (where the rocks are slightly hornfelsed), Geerabukta and Sorgfjorden, in all of which the typical Member 2 lithology is well displayed. Around Sorgfjorden the rocks are strongly disturbed; thus on Heclahuken, on the east limb of the "Gorge Valley Syncline" (Fleming and Edmonds 1941), the formation is less than 200 m. thick and the rocks are strongly sheared, while on the west limb north of Dunérbreen the apparent thickness is at least 500 m., and the beds compressed with small-scale isoclinal folding. The discrepancy is explained by incompetent folding in connection with the higher degree of metamorphism prevailing in this area.

(*b*) *The Upper Grusdievbreen Limestones*

These rocks are unique in the following respects: although pure pale grey limestones are prevalent, the rocks are characterized by diffuse

banding of pale, medium and dark grey limestones (Plate V, fig. 4); the general abundance of mottled limestones (on a false bedded pattern or due to orthogenic quartz clots), of pseudo-conglomerates, and calcite vein-breccias; the type and mode of occurrence of *Collenia*. Non-calcareous material is absent, except at the extreme base and top of the formation.

(i) Sections on Svanbergfjelet and Golitsynfjellet provide between them a fairly complete sequence which is divided as follows:—

6. 90 m. Limestones, banded, pale grey.
5. 50 m. Limestones, banded, pale grey and grey-black.
4. 150 m. Limestones, banded, pale grey.
3. 75 m. Limestones, banded, pale grey and grey-black.
2. 90 m. Limestones, banded, pale grey, with massive base.
1. ?10 m. Calcareous flags, mainly red-brown.

The thicknesses, except those of Members 1 and 6, are reliable to within 15%.

Member 1 consists of laminated red-brown calcareous siltstone flags with well developed false bedding, intra-formational pebble, flake and edgewise conglomerates and also some bluish green-weathering dolomitic shales at the base. Some of the red beds are partly or completely dolomitized. A thin section of the red flags reveals an abundance of strained quartz grains, mostly well rounded, with muscovite in small laths arranged sub-parallel to the bedding. Intermediate plagioclase occurs in occasional grains, and carbonate is predominant along certain bands. The matrix is largely of limonite or haematite, which is concentrated in short irregular streaks and is largely responsible for the lamination.

Member 2 is of rather monotonous appearance, but closer examination shows it to be variable, as is illustrated by the following detailed example from about the middle of the division on Fulmarberget:—

f. 1 m. Limestone, pale grey, structureless, rough-weathering.
e. 1 m. Limestone and dolomitic limestone, laminated, smooth, weathering grey and buff respectively.
d. 1 m. Limestone, grey, mottled.
c. 1 m. Limestone, pale grey, oolitoidal.
b. 1 m. Limestone, grey, mottled on a false bedded pattern.
a. 1 m. Limestone-siltstone, medium grey, tough, uniform.

Of the above sequence types *a.*, *d.*, and *f.* are abundant throughout the formation as a whole.

The *mottled limestones* of uncertain origin deserve special mention since they are particularly characteristic of this formation. Four main types were recognized, of which *b.*, *c.*, and *d.* above are three.

d. consists of a pale grey crystalline limestone with isolated clots composed of orthogenic quartz crystals projecting from a central cherty area. They appear in hand specimen as irregular darker grey patches a few millimetres across. In thin section the quartz prisms are seen to contain relict carbonate

rhombs in various stages of resorption. Evidently orthogenic quartz is of wide occurrence in the formation, for when various of the limestones are dissolved in acid, tiny euhedral quartz prisms are invariably present in the residue, and in the less argillaceous examples they form the bulk of it.

b. is of a curious lithology which was not seen outside the formation. Medium grey spots about ½ cm. across are arranged along curved planes of large-scale false bedding within a paler grey limestone. Individual beds may be a metre thick, and the inclination of the planes as much as 50°. The view that this may be the result of shearing is supported by the abundance of twinned, elongated crystals aligned parallel with the normal bedding. The origin of the structure is not clear, since in thin section the spots are indistinguishable from the remainder of the rock, all consisting of pure calcite of uniform crystalline structure.

A third type of mottling, and by far the most abundant, may be described as pseudo-conglomerate. Irregular, flattened areas of darker grey limestone, averaging a centimetre or two across, are distributed loosely in a paler grey rock which in thin section is of the same crystalline structure and grain size (0·1 mm.).

Finally there is mottling of quite a different kind, as in *c.* (Plate VII, fig. 3), which thin sections show to be due to oolitoids (see Section IV, (*c*).).

Prominent yellowish-weathering, partly dolomitized *Collenia* bands occur both on Fulmarberget and Golitsynfjellet. On the former there are three bands about ½ m. thick and 3 m. apart. Individual colonies are usually poorly developed (Text-fig. 2a), being of rather irregular shape and of smaller dimensions than those characteristic of the Akademikerbreen Series. The colonies pass downwards through wavy-bedded into normal banded limestones.

The remainder of the formation was examined briefly on Golitsynfjellet and Svanbergfjellet. Divisions 3 and 5 are rather similar; the grey-black limestones are usually strongly veined by white calcite which, in extreme development, gives rise to vein breccia of dyke-like form (Plate IV, fig. 2). The rocks themselves are usually mottled on a large scale and there seem to be gradations from mottled limestones to vein breccias in places, but more observations and specimens are necessary. Massive yellowish-weathering *Collenia* beds are conspicuous and sharply defined, and do not normally exceed a metre in thickness, but large flattened reef-like masses project as much as a further 2 m. into the overlying beds (commonly grey-black limestone flags) which overlap them. Locally the *Collenia* rock projects down into and veins the underlying dark limestone, with the formation of limestone breccias (Plate V, fig. 4). It is possible that the breccia shown in Plate IV, fig. 2 is also of this origin, but usually the direct connection of breccia with *Collenia* bands is not seen.

In Oxfordbreen are some prominent vertical orange-weathering dolomite-sandstone dykes about 30 cm. thick. A thin section shows the dyke rock to be a highly porous dolomite with well formed rhombs strongly stained by limonite, subordinate well-rounded grains of quartz, and a sparse limonitic matrix. There is no textural change in either dyke or wall rock towards the contact, which is very sharp. Evidently

these dykes are post-Caledonian since they show no signs of the disturbance which has caused crystal orientation and twinning in the calcite of the wall rock; however, their origin is not apparent from field observations.

Members 3 to 6. Divisions 4 and 6 are again very similar, and resemble the paler bands of Divisions 3 and 5. There is a distinctive though rather diffuse banding of pale and medium grey limestone (Plate V, fig. 4), and bedding planes are commonly marked by large stylolites up to 15 cm. in height. In Division 4 mottled limestones are particularly abundant, and in Division 6 (which is poorly exposed) some of these are on a false bedded pattern. *Collenia* bands and lenses are conspicuous, (Text-fig. 2b).

(ii) Further north few detailed stratigraphiçal observations of the formation were made, though suitable sections may be found around Dracoisen and Lomfjorden. Bordering the former, in Polarisbreen, some red-brown silty flags very like those of Division 1 were found in a narrow scree zone at approximately the same horizon, and the general lithology of the surrounding rocks is very much as in the south. On Raudberget cliff a lateral transition from the pale to the dark facies along a bedding plane was observed. In Glintdalen the character of the rocks is largely unaltered, but there also appear some brownish-weathering beds and pale cherty limestones rather similar in appearance to the less conglomeratic rocks of the Draken Conglomerates. The position of the upper boundary was again not fixed here, but what are probably the uppermost beds contain much black chert in the form of pancake-shaped concretions up to 60 cm. across and averaging 10 cm. in thickness. These are aligned along certain bedding planes in white dolomitic limestones, similar to beds noted on Svanbergfjellet.

(iii) In north Ny Friesland there is an extensive outcrop south and east of Sorgfjorden, and the following section was roughly estimated in ice-worn rocks on the east margin of Dunérbreen ice front:—

h.	30 m.+	Dolomites, grey-white, cherty; regarded as the basal beds of the Akademikerbreen Series.
g.	60 m.	Limestone flags, pale grey.
f.	60 m.	Limestones, banded, thickly bedded, weathering orange-brown and medium grey.
e.	60 m.	Limestones, cream-white, thinly bedded.
d.	½ m.	Flags, red-brown and green-grey.
c.	5 m.	Orange-brown-weathering beds.
b.	50 m.	Limestones, alternating grey and cream-white, thickly bedded.
	?200 m.	unexposed gap.
a.	?500 m.	rocks of typical Lower Grusdievbreen Limestone type.

Even allowing for incompetent folding, no detailed correlation with the Akademikerbreen section seems possible. All original sedimentary structures have been destroyed by the deformation, but the lithology as a whole is sufficiently distinctive for there to be no confusion with

any other formation. Consequently the isolated mounds of pale limestone projecting through the outwash gravels north of Dunérbreen are referred with confidence to this formation, and not to the Akademikerbreen (Ryssø) Series as suggested by Fleming and Edmonds (1941); this is also more in conformity with their structural position.

The Raudberget Breccia

An exposure on Raudberget is of special interest in that the Upper Grusdievbreen Limestones are interrupted by an extensive boulder bed (Text-fig. 2a and b and Plate IV, fig. 4) differing from any known elsewhere, which extends across an almost vertical boss-like cliff face about 150 m. high aligned at right angles to the normal strike of the Hecla Hoek. It was discovered in 1951 in the course of a rapid sledge journey by M. B. Bayly, M. H. P. Bott, and W. B. Harland, who considered a fault breccia but suggested (privately with some doubt) that it could be a land tillite with large boulders frost shattered in situ. In 1956 the author was able to spend two days there, and made the following observations.

(i) The boulders are either angular or sub-angular.
(ii) They are quite unsorted and range up to a metre across.
(iii) The boulders, though apparently of great variety, are found to consist entirely of limestones derived from the immediate strata which have been permeated by haematite in varying degrees.
(iv) No glacial striae were seen.
(v) The matrix commonly veins the boulders, and all stages can be found between this and complete shattering into angular fragments.
(vi) The matrix consists entirely of highly ferruginous limestone, and both clay and quartz are virtually absent.
(vii) The calcite of the matrix is untwinned.
(viii) Boulder beds are both interbedded with and appear to pass laterally into normal unbrecciated limestone at the top of the cliff, and may form a roughly stratified feature which covers the entire cliff as though draped over it like a curtain. This is accounted for by a sharp local swing of the strike in connection with a strike fault which follows a gully straight up the cliff. Thus the true thickness of the boulder beds may be as little as 30 m.
(ix) The boulder beds thus appear to be the fault breccia of the above-mentioned strike fault.
(x) There occurs within the opposite wall of the fault plane a prominent off-shoot from the fault, from which it is quite clear that the country rock has been penetrated and shattered by invading fluids passing up along the fault plane.
(xi) Similar breccias, though on a much smaller scale and without iron oxide impregnation, are of frequent occurrence in this particular formation in other localities.
(xii) Some strongly ferruginous brecciated limestone similar to that on Raudberget was seen in the south lateral moraine of Polarisbreen, and may possibly be represented in situ on the northern side (unvisited), and this was probably derived from the same formation. This suggests that the feature extends southwards along the strike without notable transgression.
(xiii) The ferruginous impregnation, although confined to this horizon on Raudberget, is of wide distribution elsewhere around Raudbergbreen, especially on Raudbergryggen where the quartzites and greywackes of the Veteranen Series have also been strongly hornfelsed by the nearby granite.

The weight of evidence suggests that it must be a vein breccia of unusual size and type, though points (viii), (xi), (xii) and (xiii) are not easily explained on this basis. The feature merits more detailed work, as at present it is not clear to what extent it is penecontemporaneous and to what extent it is related to the intrusion of the Chydenius (Raudbergbreen) Granite.

III. The Akademikerbreen Series

Apart from the overlying Polarisbreen tillites, this series contains the most varied rock types in the Hecla Hoek, including limestones, dolomites, sandstones, a variety of shales, cherts, and occasional iron-stones; and abundantly represented among the sedimentary structures are *Collenia* structures, penecontemporaneous carbonate conglomerates and micro-conglomerates, oolites and oolitoids.

It has been shown that they are considerably thicker than at first supposed, from the discovery of a strike fault in the original section and a small though distinct unconformity in Elbobreen within a horizon of black flaggy limestone. Although a minor one, this was the first unconformity to be demonstrated in the Hecla Hoek of Ny Friesland. The angle of discordance is about 15 degrees and the surface of unconformity is smooth. Apart from this the supposed thinning to the North is now confirmed, as is the correlation with the much thinner section of Kluftdalen ("Gorge Valley"). However, the unconformity has only been proved in one place, and since there must by inference be other minor breaks higher up in the Svanbergfjellet Beds, it is not used as a formational boundary.

Excellent exposures occur in Kluftdalen, Faksevika, Glintdalen, Raudberget, Polarisbreen, Oslobreen, Ellevepiggane and the Backlundtoppen type area, in all of which the formations can be recognized by their general lithology and sequence. The Ryssø dolomite undoubtedly corresponds with the massive *Collenia* beds which occur in both the Backlundtoppen Dolomites and the Svanbergfjellet Beds. The intervening Backlundtoppen Oolite was first interpreted in Fakesvika by Kulling as a repetition by faulting of the Lower Hunnberg Series. Massive *Collenia* bands occur undoubtedly well down into the Grusdievbreen Series.

(*a*) *The Svanbergfjellet Beds*

The type section on Svanbergfjellet divides naturally into four members:—

4. The Upper Svanbergfjellet Limestones.
3. The Svanbergfjellet *Collenia* Beds.
2. The Lower Svanbergfjellet Limestones.
1. The Svanbergfjellet Dolomites.

Although these names indicate only lithological preponderance, the boundaries are all well defined, with the exception of that below Member 1 as already explained. In central Ny Friesland, however, the base of Members 2 and 4 also becomes uncertain, though the same general sequence applies here as in the south (see Text-fig. 1). In Kluftdalen the sequence is much attenuated, in view of which the general similarity is remarkable. Detailed sequences were obtained of only certain small parts of the formation, particularly in Member 4 and across the boundary between Members 1 and 2. Complete exposures are almost limited to Akademikerbreen and Geerabukta, though fairly extensive outcrops may be found in Grusdievbreen, Polarisbreen and Glintdalen.

1. *The Svanbergfjellet Dolomites* of Akademikerbreen are mainly well bedded, grey-white-weathering, pale grey dolomites and dolomitic limestones usually containing black and white chert in the form of lenticles and nodules, but in the lower beds white chert occurs also as a fine rectangular network in which films at right angles to the bedding are as common as those parallel to it. Intra-formational conglomerate bands are subordinate except in the uppermost beds which, since they also contain a few continuous layers of black chert, greatly resemble the typical Draken Conglomerates. There are also some bands containing large ellipsoidal chert concretions flattened parallel to the bedding. Isolated *Collenia* bands, not usually exceeding a metre in thickness, occur frequently and are usually of the same type as is characteristic of the Backlundtoppen Dolomite. All the *Collenia* bands are pale, with the exception of two which occur in dark grey limestone. In these the *Collenia* is very distinctive (Text-fig. 2c and e); one band occurs at the top of the formation in the south, and the other forms a conspicuous feature above Geerabukta in the lower part of the division.

The uppermost 10 m. are of a contrasted character, the same sequence being found on Svanbergfjellet and Golitsynfjellet as follows:—

e. 1 m. Cherty flags, weathering brown.
d. 3 m. *Collenia* dolomite, massive, weathering brownish (Text-fig. 2c)
c. 5 m. Dolomitic flags, green-grey, weathering cream and green-grey, fine grained with wavy bedding planes.
b. 3 m. Paper shales, blue-black.
a. ½ m. Quartzite, white, rust-coated, laminated, fine grained.
Thick pale dolomites.

A similar sequence was noted in Polarisbreen where also the main dolomites are brownish-weathering, with bands of angular breccia a metre or so thick. Since this is not the case in Glintdalen or Geerabukta the facies must be local. It is not clear whether this division is present in Kluftdalen, since 30 m. of beds of the same lithology turn out to be limestones instead of dolomites.

2. *The Lower Svanbergfjellet Limestones* in Akademikerbreen form steep cliffs where the predominant rock type is a thickly bedded or massive grey to grey-black limestone, generally iron-stained and calcite-veined with interbedded paler layers in the upper part. The basal beds show extreme regularity of bedding (Plate V, fig. 2) which, combined with the convex weathering profile of each bed, suggested the name " Bolster Beds ". The rocks consist of thinly bedded grey-brown-weathering, medium-grey, very fine-grained dolomitic limestone with shaly partings of similar material. They form one of the distinctive marker horizons for south and central Ny Friesland. The base, at least, of the overlying dark limestones contains large (30 m.) ochre-weathering cherty concretions, and a few bands of black and white limestone-dolomite breccia.

In Polarisbreen the pattern is similar, with Bolster Beds and paler layers towards the top. The latter consist of either massive pale dolomite or yellowish-weathering coarsely porous dolomite, possibly produced by differential removal of pisoliths after dolomitization. The lithology is strikingly reminiscent of beds in the Veteranen Limestones (Wilson 1958). Bolster Beds were not recognized north of Polarisbreen, though in Geerabukta and Kluftdalen are horizons of black limestone at the appropriate level which are probably equivalent to the main part of the division.

3. *The Svanbergfjellet Collenia Beds* form a unique lithological division consisting of massive bands of *Collenia* dolomite, variously coloured by ferruginous matter, separated by knife-sharp boundaries from thinner layers of dark, mainly purple and green, siliceous paper shales. Except possibly at the very top no other rock types are present, and the contrast between the dolomites and shales is very striking. In the field the shale bands are often obscured, and in fact the whole division tends to be elusive, probably due to the prevalence of columnar jointing in the dolomites which renders them very crumbly in spite of their massive nature. Unlike the Backlundtoppen Dolomite *Collenias*, these occur as distinct beds averaging about 3 m. in thickness, separated by large scale stylolites with towers sometimes exceeding 15 cm. in height (see Section IV (*f*)).

No attempt was made to subdivide the sequence, but general traverses on Svanbergfjellet, Macdonaldryggen and above Geerabukta, together with partial sections elsewhere, indicate that the main features are unchanged throughout Ny Friesland in spite of the extreme thinning in the north, where the association of ferruginous *Collenia* dolomite and more than 10 m. of coloured paper shales is a sufficient basis for the identification of this division.

The *Collenia* themselves were scarcely studied, but are of several varieties including those illustrated in Text-fig. 2.*i*, *c*. and *d*. Some thin

bands with reef-like upper surfaces were also seen at the top of the division in central Ny Friesland, interbedded with banded dolomite-siltstones and dark grey intra-formational dolomite conglomerates.

4. *The Upper Svanbergfjellet Limestones* include a varied group of sediments of which good exposures are few. Fairly detailed sections were obtained from Akademikerbreen, and some others elsewhere were more roughly estimated (Text-fig. 1). Although there is a general similarity in the upper beds, the lower beds differ much in detail, and thus it is not certain that the lower boundary is correctly placed in the central Ny Friesland sections, since the basal pale dolomites have no obvious equivalents in the type section to the south. The generalized sequence for Akademikerbreen is as follows:—

g. Quartzite, grey, weathering brown, laminated, impure.
f. Limestone, black, oolitic and pisolitic, becoming more thinly bedded downwards and grading into:—
e. Limestone and dolomitic limestone flags in alternation with wavy bedding planes. On Macdonaldryggen dark siliceous shales appear at the base.
d. Siliceous and dolomitic flags, pale green-grey, weathering pale green-grey and orange respectively.
c. Calcareous flags, partly of fine grained limestone-siltstone, pale green and reddish, weathering yellow and red respectively.
b. As *f*.
a. As *e*.

Another probable exposure of these beds is on Løva, Alexanderbreen, visited by M. B. Bayly, from whose notes the following sequence is taken:—

l. Conglomerate, with rounded fragments up to 10 cm. diameter of limestone, cherty limestone, etc., in a matrix of fissile grey limestone.
k. Limestone, pale, with concretions, intraformational conglomerate and *Collenia*.
j. Quartzite band, weathering brown, with concretions in which the original rock is partly or entirely replaced by fine grained chert.
i. Limestone, grey, weathering yellow, with vertical streaks of recrystallized calcite, suggesting the rising of bubbles before consolidation.
h. Limestone, black, massive, with spheroidal structures.
g. Quartzite band, weathering brown, with pebbles and iron-stained concretions.

(*g*.–*k*. 20 m.).

f. 15 m. *Collenia* band with reef development.
—undulating erosion surface—
c. 15 m. Limestone, black, massive, with top part containing pisolite, desiccation cracks, flake conglomerate, etc.
d. 30 m. Limestone, black, weathering yellow, with occasional massive bands. Much calcite veining.
c. Limestone bands, weathering gold, projecting from scree.
b. ?Green and purple shales
a. ?Limestones, white and grey } unvisited.

(*a*.–*c*. 60 m.).

There seems little doubt that beds *d.* to *h.* represent the Upper Svanbergfjellet Limestones, and beds *i.* to *l.* the basal Draken Conglomerates.

The sections in Polarisbreen and Chydeniusbreen are either incomplete or partly scree-covered, and the only other satisfactory exposure is above Geerabukta where the following sequence was paced out:—

k. Mainly as Akademikerbreen *f.* and *e.*, but with a variety of conglomeratic layers containing ovoid pebbles of black chert, pale-weathering dolomite and black limestone in a matrix of grey limestone. Some bands have strongly desiccated surfaces.

j. Calcareous siltstone flags, pale brick-red and pale green-grey, with desiccation cracks and strong false bedding.

—sharp boundary—

i. ½ m. Limestone flake-conglomerate, partly dolomitized, banded cream-orange and grey-weathering, capped by 2 cm. of pale green porcellanous limestone with a rough, nobbly upper surface.

h. 2 m. Limestone with edgewise conglomerate, partly laminated, grey-white, weathering yellow-brown: clasts replaced by pale chert.

g. ?2 m. As *h.* but weathering reddish and more flaggy.

f. 1½ m. As *h.* but with ochre-weathering surfaces.

e. 1 m. *Collenia* limestone, massive, weathering yellow-brown, with a well developed reef-like upper surface. The limestone is partly dolomitized and silicified, and the associated beds are strongly false bedded.

d. ½ m. Haematitic flake-conglomerate, with sharp lower boundary.

c. ?2 m. Dolomite-siltstone, weathering cream-orange, smooth, strongly false bedded with large flake conglomerate layers.

b. 1 m. Shales, grey.

a. 2 m. Dolomite, cream-brown-weathering, rough, with very large wave markings of 2 m. wave length, otherwise similar to *c.*

The Svanbergfjellet *Collenia* Beds.

The main differences between this section and those of Akademikerbreen are the abundance of conglomeratic layers and the differing lithology of beds *a.* to *e.* which, however, could equally well belong to the Svanbergfjellet *Collenia* Beds.

(*b*) *The Draken Conglomerates*

(i) *Stratigraphy.*—The formation is especially well exposed in Backlundtoppen (Plate V, fig. 1), Polarisbreen, Raudberget and Geerabukta and also partly in Draken, Mjølfjella and Kvitbreen, in all of which it forms steep cliffs with great distinctive slab-like beds. The lower beds are massive in the south, but the upper more distinctive part is conspicuously, though thickly bedded, and divided at intervals by dark grey and black bands which are variously of calcareous shale, black chert or grey oolitic limestone. The sequence on Backlundtoppen is as follows:—

m. 150 m. Undifferentiated cherty dolomites, pale grey, weathering grey-white, mainly thickly bedded with abundant intraformational conglomerates of various types, often differentially silicified

by black or white chert and with some continuous bands of black chert up to a metre thick with residual sedimentary structures. There also occur at intervals thin bands of *Collenia* dolomite, dark grey shale and oolitic or pisolitic rock.

l. ½ m. *Collenia* dolomite, weathering grey-white (Text-fig. 2(*i*)).
k. 5 m. Oolite, medium grey.
j. 10 m. Dolomites as in *m.*, but massive.
i. 25 m. Dolomites as in *m.*, bedded.
h. 8 m. Pisolite, massive, with white-weathering pisoliths in a medium grey-weathering matrix.
g. 42 m. Dolomites as in *m.*, though generally paler and less cherty.
f. 10 cm. Dolomite-sandstone, brownish-weathering, strongly false-bedded.
e. 27 m. Quartzite, laminated, weathering brown.
d. Siliceous flags, black, iron-stained.
c. 50 m. Dolomites, massive bedded, weathering grey-white with oolites, flake conglomerates and stylolites.
b. Dolomite, medium-pale grey, with abundant gravel-stones.
a. 3 m. *Collenia* dolomite, dark grey, weathering yellow-brown. A prominent resistant band.

Total 300 m.

The uppermost beds were traversed on Draken, as follows:—

g. 5 m. Dolomite, weathering pale, thickly bedded.
f. 8 m. Dolomite, pale grey, massive.
e. 9 m. Limestone and dolomitic limestone flags, dark grey, weathering grey and cream respectively (much as in the Lower Grusdievbreen Limestones Division 2).
d. ½ m. *Collenia* dolomite, massive, partly silicified.
c. ½ m. *Collenia* limestone, massive, not silicified but differentially dolomitized, producing a striped grey and cream rock.
b. 3½ m. Limestone, grey-white, massive. The lower half contains outlines of spherical structures 1½ m. in diameter resembling concretions.
a. Thick conglomeratic dolomites, only partly exposed on Draken.

The unusual rock type *c.* was found at an equivalent level on Ellevepiggane and in Polarisbreen. The structure in *b.* was not noted elsewhere. In Polarisbreen and Chydenuisbreen the sequence is similar to that of Backlundtoppen, though the basal *Collenia* band was not seen and the lower massive beds are much thinner. Also on Raudberget there occurs a narrow horizon of red flags above the latter. In Geerabukta and Kluftdalen the massive beds are absent.

(ii) *Petrography.*—The conglomerates and gravelstones are all apparently intra-formational and are of four main types :—

i. Gravelstones as in Plate VII, fig. 1. Ovoid and somewhat distorted clasts of varying grain size and argillaceous content are tightly packed in a sparse matrix of clearer and more coarse grained dolomite. In some cases there appears to be a transition from clast to matrix so that they are hard to distinguish. This variety is extremely abundant at all levels, and several thin sections show identical features.

ii. Pebble conglomerates with ovoid pebbles of dolomite or chert

scattered in a structureless dolomite matrix which is usually clearer and more coarse grained.

iii. Flake conglomerates, commonly with flakes of black chert in a structureless dolomite matrix. The flakes are sharply defined and almost certainly local, derived from a thin layer of black chert (Plate VII, fig. 4).

iv. Occasional bands of gravelstone up to 1½ m. thick resembling pellet dolomites (Plate VI, fig. 5), with white-weathering "pellets" in a medium grey matrix. Partly rounded and undistorted clasts of very fine grained dolomite are closely packed in a dolomite-sandstone matrix formed of angular grains of similar materials. Although clasts grade down into matrix material, the sediment is relatively well graded in maximum and minimum grain sizes. There is, as is usual in all the above conglomerates, a considerable variety of argillaceous impurity among the clasts, though the bed cannot be termed polygenetic.

Plate VI, fig. 1 shows a curiously mixed sediment containing scattered ooliths in association with flakes and shadow structures, and also aggregates of small clasts, the distribution of which is hard to account for except by supposing that they were originally contained by a marginal zone which has subsequently been removed. Plate VI, fig. 3 shows an enlargement of one of these, seen to be composed of ovoid grains with a tufa-like outer margin which shows a faint cross under crossed Nicols. This margin has been partly removed around the outer margin of the aggregate. These structures always appear distinct from ooliths or oolitoids, never having more than one zone of accretion.

The *Collenia* bands are not more than a few metres thick and are mainly of Backlundtoppen Dolomite type, with regular straight-sided colonies and strongly convex zones, but reef-like upper surfaces are very common and are overlapped by the overlying sediments. The reefs may be over 3 m. in diameter but are not very convex.

Chert is abundant throughout the formation, both as differential replacement of sedimentary structures and as continuous layers in which the matrix is also affected. All thin sections showed that relicts of original conglomeratic structure were preserved, either as zones of residual dolomite rhombs or as a faint brownish stain in the chert. In some conglomerates (as in type *c*. above) the chert evidently pre-dates the conglomerate, but in other cases there is no sharp distinction between partly silicified clasts and partly silicified matrix, and the replacement was probably subsequent. This is almost certainly true of the continuous black chert layers, the lower boundaries of which are often transitional as seen in thin section (Plate VI, fig. 2). On the other hand the upper boundaries are frequently erosion surfaces, and the sharp chert-dolomite contacts in these cases suggest that the chert pre-dated the overlying carbonate (which may have been limestone at the time),

as it is otherwise most unlikely that the silicification would have such respect for the ill-defined micro-bedding planes characteristic of this formation. Again, the black chert is invariably arranged parallel with the bedding planes and does not transect them to any depth, as might be expected were the replacement post-formational. Thus in the opinion of the author all the black chert is intraformational replacement of carbonate, much of it being almost penecontemporaneous. It was not determined to what depth silicification was able to affect beds already laid down.

Although the white chert was less conspicuous it was frequently seen to occur as cross-cutting veinlets, probably of later date than the black chert.

(c) *The Backlundtoppen Oolites*

(i) *Stratigraphy.*—The following general sequence applies throughout Ny Friesland with remarkable constancy, and may be divided into three members.

Member 1: 100–300 m. mainly of thickly bedded, dark grey to black, cherty limestones with numerous oolitic and pisolitic bands, subordinate and less conspicuous flake conglomerates, pseudo-conglomerates and mottled limestones, and intercalations of dark grey calcareous shale. There also occur a few pale-weathering dolomitic bands from ½ m. to 4m . thick, consisting variously of rock resembling Draken Conglomerates, Backlundtoppen Dolomite, or a pale grey-weathering oolite. The base of the division is invariably shaly, partly siliceous and brownish-weathering.

Member 2: Medium grey pale-weathering dolomite is coarsely stratified, and though sometimes containing *Collenia* and much pale chert it is commonly featureless. The upper boundary is usually sharp and the lower boundary transitional.

Member 3 is mainly of dark grey dolomitic limestone, commonly rusty-weathering and very irregularly bedded. Massive beds of black limestone alternate with thin layers and lenticles of paler pisolitic and oolitic dolomite, partly or completely silicified—the " coloured oolitic horizon ". Calcite veining is prominent.

(ii) *Petrography of the Oolites.*—By far the most common sedimentary structures are oolites and pisolites which are of two main types, characteristic of Members 1 and 3 respectively.

The principal variety of Member 1 seems to be confined to the dark grey or black limestone-siltstone facies, which is usually partly replaced by black chert. Its characteristics are as follows:—

(i) Thin sections show that the ooliths are clearer and more coarsely cystalline than the matrix, and unlike the latter are usually tinted red-brown by finely divided haematite, which is not conspicuous in the muddy grey matrix.

(ii) The ooliths are formed, particularly in their marginal zones, of arcuate calcite crystals arranged concentrically by zones, but without corresponding optical orientation (Plate VI, fig. 6). The central areas are commonly of irregularly crystalline calcite without nuclei, but arcuate zones may alternate with irregular ones and many small ooliths are entirely irregular. Larger calcite rhombs occasionally cut across the regular zones and transect zonal boundaries, indicating that recrystallization has taken place.

(iii) The arcuate crystals are traversed by a very fine and closely spaced radial structure from which the present ooliths have evidently recrystallized.

(iv) The matrix is always distinct in appearance and sharply defined from the ooliths, and invariably consists of uniform, equigranular calcite mudstone. There is frequently a faint and diffuse zoning of unknown origin concentric with nearby oolith margins.

(v) Many of the rocks are partly or wholly silicified, the ooliths being preferentially affected. The silica replacing the ooliths is always chert, but that replacing the matrix may be chalcedony as well as chert. In either case the argillaceous impurities are largely expelled, though much more effectively by the chalcedony which, however, still retains the faint zoning mentioned above. The iron ores are either expelled from ooliths or arranged along radial or concentric zones, commonly as granular magnetite.

(vi) Flattened and distorted ooliths are abundant and are not the result of regional shearing.

(vii) Composite ooliths are common, the interspaces being filled with typical matrix mudstone.

The upper part of Member 3, which everywhere contains a great variety of coloured pisolitic rocks not paralleled at other levels, is termed the " Coloured Oolite Horizon ". Examples were found of all stages of silicification from dolomite, with a variety of weathering features ranging from removal of ooliths to partial removal of the matrix, as well as differential colouring of ooliths and matrix in shades of pink, red, brown, green, grey, yellow, and white. On Backlundtoppen, whence most of the specimens came, the pisolites occur typically as clots and lenticles within the dark grey limestone forming the remainder of the division.

A few thin sections show that any original crystalline structure is completely destroyed, and zoning is often rudimentary or absent (Plate VI, fig. 6). Where the matrix is partly removed by weathering, the oolith margins are lined by radially arranged euhedral dolomite rhombs: where the pisoliths are removed the matrix is of silica and the rock resembles a pumice. In the purely dolomitic varieties the crystalline texture inside and outside the pisoliths is identical, though the grain size is often smaller within the pisoliths: the zoning is indicated either by clay impurity or by scattered pyrites or haematite. At the centre of a few pisoliths there are some obscure, steeply spiral, branching or irregular structures.

(*d*) *The Backlundtoppen Dolomite*

This is perhaps the most conspicuous and easily identified formation of the Hecla Hoek, and doubtless corresponds with the typical Ryssø Formation of Hinlopenstretet. It includes also a horizon of dark

shales and pale dolomite above the main mass, and the general sequence, which can be followed as far north as Chydeniusbreen without noticeable change, is as follows (see Plate V, fig. 1):—

3. 6–25 m. Dolomite, massive, pale grey.
2. 18–20 m. Shales, dark, quartzose and multi-coloured.
1. 150–250 m. Dolomites, massive, pale grey.

Member 1 which forms the main mass can be recognized everywhere on its combined lithology and thickness; thus the yellowish or pinkish-white colour of weathering, the almost complete absence of stratification, the very coarse and irregular mode of weathering on cliff faces and the absence of other rock types are features which are not paralleled in any other formation, except as isolated bands less than 10 m. thick. Most of the beds, especially in the upper part, are composed of massive or separated *Collenia* colonies often intimately associated with dolomite breccias, oolitic and oolitoidal beds (see Section (IV*c*). These latter do not form conspicuous bands, but judging by a few thin sections the structures may be widespread. Towards the top the breccias are abundant, with angular fragments of pale grey dolomite, often shattered *Collenia*, in a pale pinkish and coarsely crystalline dolomite matrix. The latter has the appearance of carbonate veining, and the rocks may be vein breccias formed in a manner similar to those associated with *Collenia* bands in the Upper Grusdievbreen Limestones.

The *Collenia* structures are typically large, with diameters from 10 to 20 cm. and heights of up to a metre. Colonies may be separate but are more usually massive, as depicted in Text-fig. 2(i). The lamination is typically rather fine (3 to 5 per centimetre), and, as in most of the *Collenia* structures in the Hecla Hoek, is produced by argillaceous layers of finer grain than the main rock but otherwise of no essentially different character. Replacement by chert is usually in the form of fine permeation, preferentially affecting the argillaceous zones. In the lower part of Member 1, where silification is less evident, *Collenia* is also less conspicuous and may have escaped notice.

The shaly beds of Member 2 are sharply defined from the dolomites above and beneath and are lithologically in marked contrast with them. The overall thickness seems to be remarkably constant, and the following subdivisions can be followed, with exceptions, from Chansjinfjellet to Polarisbreen:—

d. ½ m. Quartzite, pure, fine grained, weathering orange.
c. 1–2 m. Paper shales, purple, quartzose and non-calcareous.
b. 10 m. Paper shales, green, quartzose and non-calcareous.
a. 5–10 m. Mixed beds including green-white quartzitic flags, orange-weathering dolomitic flags and black shales; beds in general weathering dark.

Subdivisions *c.* and *d.* are absent from Raudberget, and further north this and the overlying Member 3 were not identified.

Member 3 is rather similar in appearance to Member 1, but differs in its coarse stratification, and contains undulose structures up to a metre across, much larger than any known *Collenia*. These structures are convex downwards, thus differing fundamentally from the genus *Collenia*. The lamination forming the structures is not continuous as in *Collenia*, but is produced by irregular flattened areas of different crystalline texture or impurity content, sometimes differentially silicified, arranged concentrically *en echelon*. Within these areas dolomite crystals are often oriented at right angles to the lamination. The cherty types, which are perhaps most typical, weather into extremely rough, coarsely porous rock, (see Text-fig. 2(h)).

Members 2 and 3 are well seen on Backlundtoppen and eastern Ellevepiggane, and in Oslobreen, Polarisbrecn and lower Chydeniusbreen. Exposures in Glintdalen and Hingstsletta are too broken, and those of Kluftdalen too poorly known, for correlation to be extended to these areas, though Member 2 may be present in the latter. There possibly, the appearance of the rocks was changed by the strong compression which induced cleavage in the overlying tilites. At any rate there can be no doubt of the presence of Member 1 in Kluftdalen. Thus the suggestion by Fleming and Edmonds (1941) that the equivalent Ryssø Dolomite is absent from this section is incorrect.

IV. Discussion

(*a*) *Cyclical Sedimentation*

Small scale repetitions of facies were not sufficiently studied, and broader cycles are difficult to recognize owing to the virtual absence of arenaceous sediments and the lack of lithological variety in all formations except the Svanbergfjellet Beds. In the Upper Svanbergfjellet Limestones, however, there occurs a recognizable facies sequence as follows:—

e. Black, thickly bedded limestones, sometimes oolitic.
d. Cream and black-weathering, banded dolomitic limestones and limestones, usually flaggy.
c. Mainly pale green-grey, soft dolomitic flags.
b. Yellow and red-weathering, laminated calcareous flags.
a. Locally a thin band of impure quartzite, slightly ferruginous.

Allowing for the virtual absence of arenaceous sediments, the agreement with the carbonate cycle already demonstrated in the Lower Middle Hecla Hoek (Wilson 1958) is exact. This is the only formation in which carbonates, coloured flags and quartzite are associated together.

The *Collenia* beds are distributed among all facies in what seems to be a quite random manner and unrelated to any facies cycle, and in fact it is these beds which delimit the above cycle, top and bottom.

(*b*) *Conditions of Deposition*

Intra-formational conglomerates, oolites, oolitoids, desiccation cracks, false bedding, ripple marks and *Collenia* beds are all regarded as evidence of shelf or shallow water sedimentation, and moreover desiccation cracks, flake conglomerates and *Collenia* suggest very shallow or intertidal waters: it seems most unlikely that any of the rocks were laid down in deep water. Regarding the evidence of the strata as a whole (Table I), minor local unconformities with overstep were seen at two levels, one within the Lower Svanbergfjellet Limestones of Elbobreen, where the strata wedge out at an angle of 15 degrees in a westerly direction, and the other at the base of the Backlundtoppen Dolomite of Oslobreen and Raudberget, but since in neither case are the underlying beds completely cut out anywhere, these breaks cannot be regarded as true unconformities. In fact all formations maintain fairly uniform lithologies and thicknesses throughout Ny Friesland except the Svanbergfjellet Beds and Draken Conglomerates, which show a pronounced thinning northwards (in Kluftdalen possibly partly tectonic), from which it might be inferred that during these times there existed an axis of subsidence with an east-west component. However, within these two formations there are a number of thin marker bands which can be traced for at least 70 km. from south to north and 15 km. from east to west, and this implies a remarkable tectonic quiescence during which subsidence was gradual, though possibly intermittent, without warping of any knd. Regarding the origin of the material, it is likely that oolites, oolitoids and *Collenia* structures were formed in situ since they probably involved the extraction of lime from solution, either by chemical precipitation or by organic agency; and although it might be argued that the remainder of the rocks could have come as detrital material from a source of purely carbonate rocks, such a hypothesis would create more problems than it solved, for two reasons. Firstly, the waters of deposition would need to be saturated in lime in any case, otherwise the detritus would at least show signs of marked re-solution, which are quite lacking; and secondly, source rocks of the necessary thickness, purity and age are quite unknown. It seems most likely, then, that the lime was derived entirely from solution and that land was either very distant or low-lying.

(*c*) *Oolitoids*

The term oolitoid (? introduced by Nordenskiöld) refers to certain sedimentary structures which, though sharing the properties of both

oolites and penecontemporaneous conglomerates, cannot rightly be classified with either, there being no intermediate types. Their properties as seen in thin section (see Plate VII, figs. 2, 3, and 6) are as follows:—

They are built up by accretion in the form of highly irregular closed zones of a type identical to those of ooliths, but generally complex. The nuclei are existing fragments or grains of any size or shape, the structures seeming capable of adsorbing much clastic matter. Inner zones are commonly abruptly truncated by outer zones, indicating that the structures were periodically broken up.

Oolitoids sometimes tend to merge with the matrix, but not to such a marked extent as in some flake conglomerates. In shape and composition they resemble the clasts of gravestones rather than ooliths, but they commonly bear an outer zone of radially arranged carbonate crystals, as do many ooliths.

The distribution of oolitoidal rocks is not known since usually they are only revealed as such after sectioning, but they may well occur at all levels, particularly in the Upper Grusdievbreen Limestones and the Backlundtoppen Dolomite. They were not noted in the dark limestone facies.

(*d*) *Carbonate Veining*

Many of the carbonate rocks are conspicuously veined, though this shows up more clearly in the dark limestone facies. In extreme development considerable thicknesses are converted into vein breccias, which may subsequently be modified so that the boundaries between clasts and veins become diffuse, giving shadow breccias. Such rocks are hard to distinguish in the field from pseudo-conglomerates which are due to some form of differential alteration without intrusion or shattering.

The age of the veining is variable; for example, much of that in the Grusdievbreen Series is of a type which suggests that the rocks were not fully consolidated at the time, and the downward projecting veining from the base of *Collenia* bands was also probably almost penecontemporaneous. On the other hand, there occur vein breccias of dyke-like form which in one instance traverse a vertical thickness of over 30 m. of strata, and these may have been of any subsequent age. The Raudberget vein breccia, again, must have been post-depositional and possibly even post-Caledonian.

(*e*) *Stylolites*

These are abundant at all levels within the carbonate rocks, and in some formations (the Upper Grusdievbreen Limestones and the Svanbergfjellet *Collenia* Beds) are responsible for much of the pseudo-

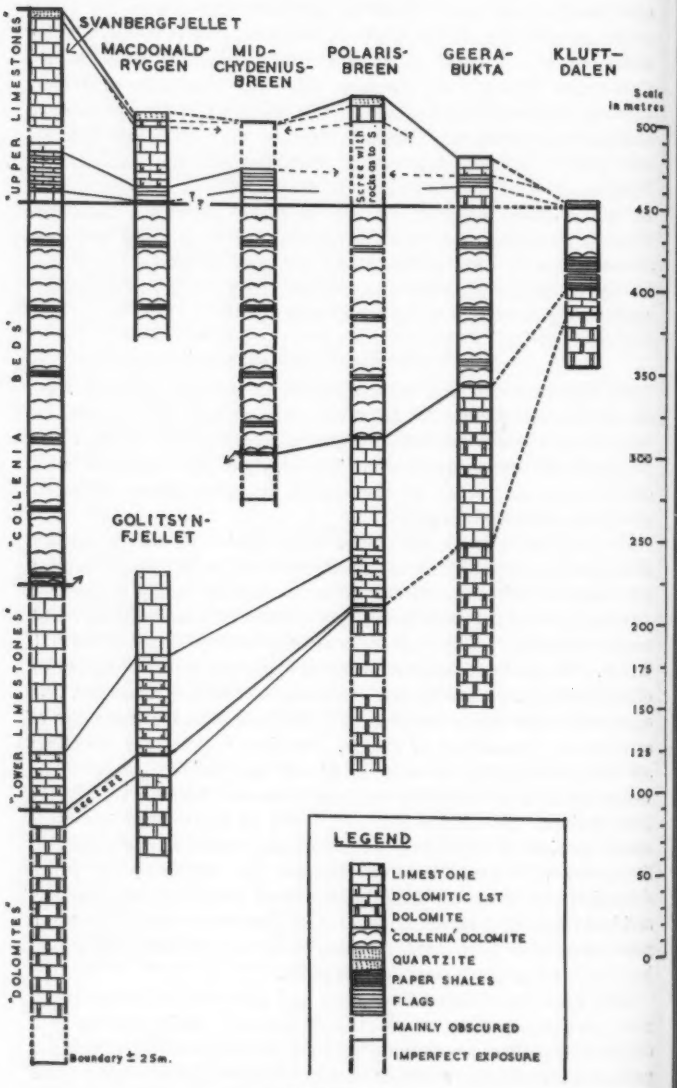
stratification with which they lie parallel. The towers are either perpendicular or at a high angle to the bedding, even in steeply dipping beds, and this is taken as proof that the structures pre-dated the Caledonian folding. At the same time they were seen to transect *Collenia* structures (e.g. Section III(a)3), sedimentary lamination, and oolites, both above and beneath, and to bear no spatial relation with them at all; they must, therefore, be post-depositional. Whether or not they were also post-diagenetic was not determined. In many cases, especially in the laminated carbonate beds and *Collenia* structures, the removal by solution of a small thickness of sediment is clearly indicated, and this is further supported by the invariable presence of the usual irregular film of insolubles representing impurities in the parent rock.

(*f*) *Recrystallization and diagenesis*

All the dolomites and almost all the limestones sectioned show clearly that recrystallization has taken place, since crystals traverse the boundaries of clasts and ooliths and in general bear no relation to the included sedimentary structures. The only possible exception is the calcite-mudstone matrix of the typical Backlundtoppen Oolite, in which the evidence is negative.

Dolomitization is almost always accompanied by the destruction of all original crystalline arrangement, both optical and morphological, an exception being the narrow zones of radially arranged dolomite crystals surrounding some dolomitized ooliths and oolitoids, in which traces of optical orientation are sometimes faintly visible (Plate VII, fig. 2). No evidence was found which suggested primary deposition of dolomite (i.e. the perfect preservation of sedimentary structures with appropriate crystalline textures), but there are abundant examples of subsequent replacement of calcite. Original sedimentary structures are preserved only by the presence of clay (unresolved), iron ores and variations of grain size: the more argillaceous rocks are usually the finer grained. A considerable proportion of the bedded carbonate rocks consists of laminated limestone and dolomite (or dolomitic limestone) with from 5 to 15 laminae per centimetre, in which dolomitization probably took place almost penecontemporaneously and in a rhythmic manner. In other cases the process could have taken place at a much later stage, affecting rocks to a considerable depth, but observations are lacking on this point.

Silification has affected limestones and dolomites of many types, more particularly those of the Akademikerbreen Series, and as with dolomitization has acted selectively on the sedimentary structures, preferentially affecting clasts or matrix. In some flake conglomerates replacement is almost penecontemporaneous (Section III(*c*)), and this


SVANBERGFJELLET
MACDONALD-RYGGEN
MID-CHYDENIUS-BREEN
POLARIS-BREEN
GEERA-BUKTA
KLUFT-DALEN
Scale in metres
500
450
400
350
300
250
200
175
150
125
100
50
0
"UPPER LIMESTONES"
"COLLENIA BEDS"
"LOWER LIMESTONES"
"DOLOMITES"
Scree with rocks up to 5
GOLITSYN-FJELLET
sea level
Boundary ± 25m.
LEGEND
LIMESTONE
DOLOMITIC LST
DOLOMITE
'COLLENIA' DOLOMITE
QUARTZITE
PAPER SHALES
FLAGS
MAINLY OBSCURED
IMPERFECT EXPOSURE

TEXT-FIG. 1.

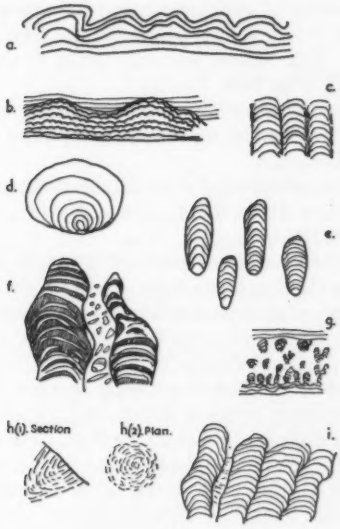
may also be true for the silica-carbonate laminated ooliths and the chert-layered *Collenia* structures. The fact that chert horizons more than a metre thick are very rare suggests that the process did not affect rocks to a depth greater than this, and that it probably only took place in rocks not yet consolidated.

Text-fig. 1.—Sections of the Svanbergfjellet Beds of Ny Friesland. The fractions indicate the maximum probable error in the thicknesses due to all causes.

Text-fig. 2.—*Collenia* and related structures from the Upper Middle Hecla Hoek rocks of Ny Friesland.

Explanation.—The nature and stratigraphical distribution of the types illustrated are as follows:—

a. Structures poorly developed with colonies characteristically small (less than 8 cm. across) and vertical partitioning almost absent. Colonies



TEXT-FIG. 2.

commonly pass both upwards and downwards into normal banded limestones. Individual *Collenia* bands average ½ m. in thickness, and the type seems to be confined to the pale carbonate facies of the Grusdievbreen Series.

b. A reef-like development confined to the pale dolomitic beds. Beds are commonly a metre thick, but the broad flattened domes may project a further metre into overlying beds. Individual colonies tend to be more complex at the base of reefs as type *g.* below, but simpler in arched domes, as type *a.* It was found between the Upper Grusdievbreen Limestones and the Draken Conglomerates.

c. This type occurs widely throughout the pale dolomitic beds of the Akademikerbreen Series and also possibly in the top division of the Upper Grusdievbreen Limestones. It is commonly developed massively in thick horizons with colonies often measuring 15 × 50 cm. A rather similar though more distinctive structure is found in the dark grey dolomitic limestone facies as a single band within the Svanbergfjellet Dolomites. Colonies are tall, straight-sided and regularly spaced, and attain dimensions of 30 × 120 cm: laminae are very smooth, strongly convex or conoidal and regularly spaced.

d. An unusual type, often more complex and irregular than illustrated, consisting of strongly silicified ferruginous dolomite of a peculiar pinkish brown colour. Colonies are separate and average 30 cm. across. This variety was only noted in the middle of the Svanbergfjellet *Collenia* Beds.

e. This type resembles the dark grey variety of *c.* except that colonies are separate. It was seen locally only in a 3 m. band at the top of the Svanbergfjellet Dolomites.

f. Differential dolomitization of laminae in a rather striking " black and white " variety found locally at the top of the Draken Conglomerates (6 cm. high).

g. Complex structures, usually difficult to make out, in the pale carbonates of the Grusdievbreen Series. Colonies are not usually more than 10 cm. across and are scattered like unstable cumulus tops: usually associated with types *a.* and *b*, but may be of wider distribution.

h. The " Dartboard Dolomite ": (1) section; (2) plan, of an unusually well developed and symmetrical structure from Backlundtoppen. The diameter is about 40 cm. and the convexity is steep downwards, thus differing fundamentally from *Collenia.* It has only been found in Member 3 of the Backlundtoppen Dolomite, but it may be present also in Member 1 which was never traversed in detail.

i. This may be the most common variety of all, and may form much of the Svanbergfjellet *Collenia* Beds and the Backlundtoppen Dolomite. It was always found as dolomite, either pale or brown weathering, which may be silicified in varying degrees. It is uncertain in what proportion colonies are massive or separate; in the latter case interstitial breccias may be present. Colonies commonly measure 15 × 60 cm.

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EXPLANATION OF PLATES

PLATE IV

FIG. 1.—A vertical ½ m. bed of pseudo-conglomerate from near the base of the Lower Grusdievbreen Limestones above Geerabukta. The beds young to the right.

FIG. 2.—A calcite vein-breccia of dyke-like form within Member 5 of the Upper Grusdievbreen Limestones of Svanbergfjellet. Maximum width of "dyke" about 2 m.

FIG. 3.—Typical dolomitized flake-conglomerate (× 1½) from the Akademikerbreen Series (Upper Svanbergfjellet Limestones of Geerabukta), as described in Section III(*i*). The order of bedding is not known.

FIG. 4.—The Raudberget Breccia (Upper Grusdievbreen Limestones). See Section II. The largest boulders are about ½ m. across.

PLATE V

FIG. 1.—Backlundtoppen from Golitsynfjellet, showing the Draken Conglomerates (DC), Backlundtoppen Oolites (BO, Members 1–3) and Backlundtoppen Dolomite (BD, Members 1–3). The outcrop is disturbed by dip-slip faults.

FIG. 2.—The "Bolster Beds" on Golitsynfjellet: about 6 m. seen.

FIG. 3.—The east face of Svanbergfjellet showing the Svanbergfjellet Beds (SB, Members 1–4), and the Upper Grusdievbreen Limestones (UGL, Members 5–6). The section traversed is indicated by the dotted line. The height of the cliff is about 600 m.

FIG. 4.—Members 3 and 4 of the Upper Grusdievbreen Limestones of Golitsynfjellet, showing about 100 m. of strata. There are numerous small dip-slip faults.

Plate VI

Fig. 1.—Thin section of a less common variety of Draken Conglomerates from Oxfordbreen. The various sedimentary structures are mainly of dolomite, and the matrix of limestone (× 3).

Fig. 2.—Banding in Draken Conglomerates (× 2).

Fig. 3.—Aggregate of small bodies with tufa-like margins from Draken Conglomerates (× 5).

Fig. 4.—Well developed pisolith from the Backlundtoppen Oolites (division 1) (× 10).

Fig. 5.—Gravelstone resembling pellet dolomite, from the Draken Conglomerates of Oxfordbreen (× 4).

Fig. 6.—Oolite from the Coloured Oolite Horizon of Backlundtoppen. The rock is entirely dolomitized, and most original structures have been destroyed (× 4).

Plate VII

Fig. 1.—Typical dolomite gravelstone composing much of the Draken Conglomerates (× 5).

Fig. 2.—Oolitoidal rock from the Akademikerbreen Series. Note the radiate dolomite crystals surrounding the structures and the abundance of orthogenic quartz (× 8).

Fig. 3.—Oolitoidal rock from Member 2 of the Upper Grusdievbreen Limestones of Fulmarbeget (× 3).

Fig. 4.—Flake conglomerate from the Draken Conglomerates of Polarisbreen. Black chert flakes in white dolomite matrix (× 2).

Fig. 5.—Limestone flake conglomerate from Member 4a of the Lower Grusdievbreen Limestones of Fulmarberget (× 10).

Fig. 6.—Oolitoidal dolomite from the Backlundtoppen Dolomite of Mjølfjella Oxfordbreen (× 2).

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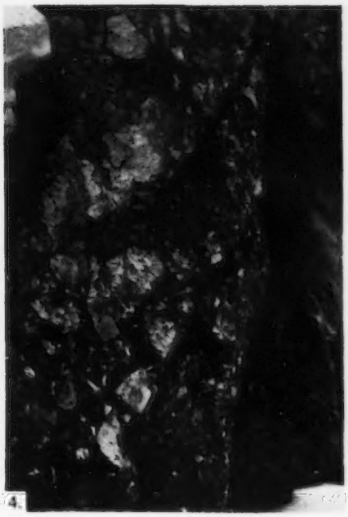

4.

UPPER MIDDLE HELLA HOEK ROCKS

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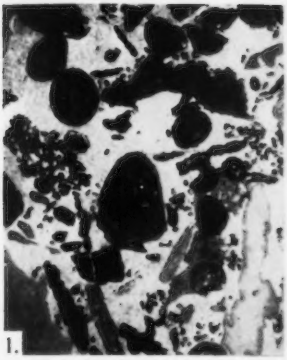
UPPER MIDDLE HELLA HOEK ROCKS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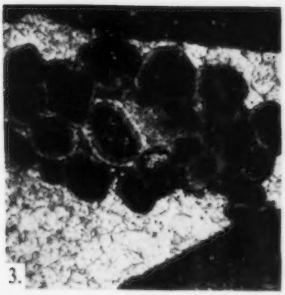
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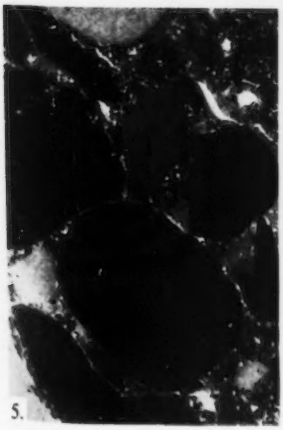
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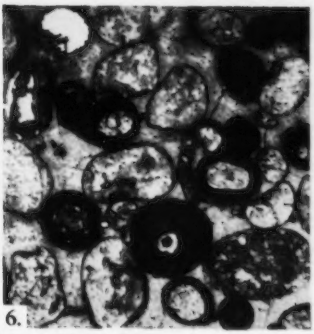
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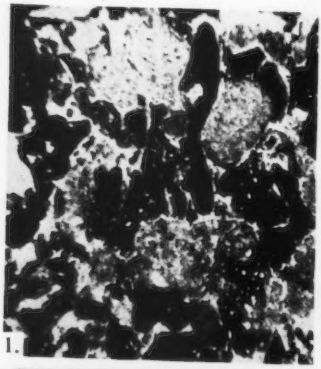
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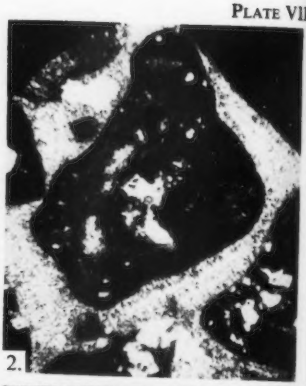
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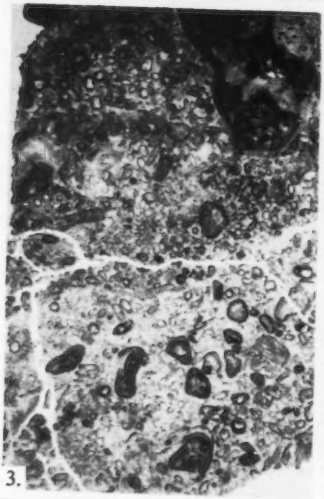
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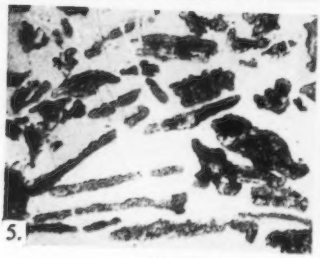
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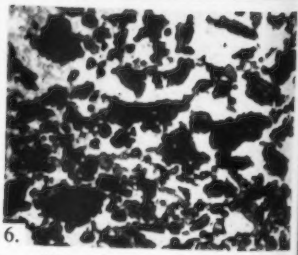
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UPPER MIDDLE HELLA HOEK ROCKS

Early Slope Development in an Expanding Stream System

By Carol S. Carter and Richard J. Chorley

Abstract

A study of the morphometry of an ephemeral stream system which is incising itself into a terrace of the Farmington River, Connecticut, supports the regional relationship between stream-channel gradient and valley-side slope angle suggested by Strahler (1950). The continuing areal extension of the system allows stream order to be employed here as a relative measure of the passage of time. Thus it is possible to recognize that steepening and parallel development of a given valley-side slope may have their place in a sequence of early stage erosional development characterized by headward channel extension and ramification.

Introduction

The problem of the development of erosional slopes through time has long formed one of the prime foci of interest for fluvial geomorphologists. Its general solution has been hampered by the complexity of the processes operating on a given slope; by the assumed variety of slope development under the influence of differing structural, lithological, climatic, and vegetative conditions; by the difficulty of observing the actual retreat of individual slopes; and, not least, by the narrow views of evolution and development stemming from the doctrinaire theories of W. M. Davis and W. Penck. The methods which have been adopted to introduce the time element into studies of slope development include: the theoretical assumption of time sequences based on inadequate considerations of process, as developed by Davis and Penck; the use of mathematical models (Bakker and Le Heux, 1946) and of scale models (Wurm, 1935–36); the successive measurement of individual slope profiles which are retreating with abnormal rapidity (Schumm, 1956); the association of sequential slope forms with successive stages in the activity of the basal streams (Strahler, 1950); and the assumption that within a single landscape are represented slope profiles of differing relative age which can be arranged in a time sequence (Savigear, 1952). The present paper presents an attempt to employ the reasoning of this last mentioned method.

The Drainage Basin

East of Lighthouse Hill, Poquonock, Connecticut, U.S.A. (Windsor Locks quadrangle), a small ephemeral stream system, here designated Lighthouse Hollow, is actively back-cutting into a terrace of the Farmington River. The terrace, the mean elevation of which is about 130 feet, is composed of very well-sorted sediments giving analysis

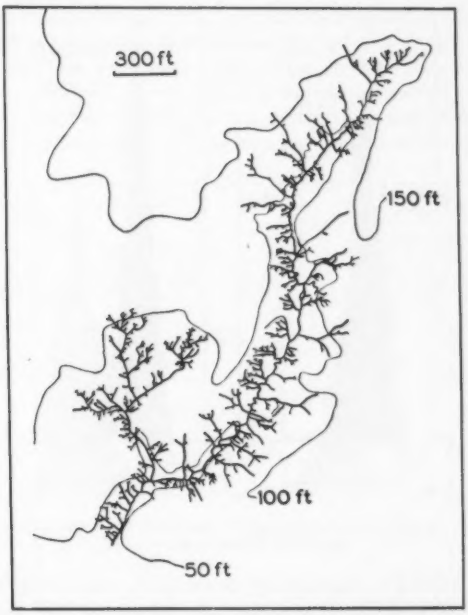
of 42 per cent fine sand and 58 per cent coarse silt sizes. Hazen's " Effective Size " (D_{10}) is ·033 mm., the range of grain size (C_r) is 1, and from Burmister's graph (1951, Fig. 13·2) it was estimated that the coefficient of permeability is approximately ·005 cm. per second at 50 per cent relative density. This high percentage of fine fractions allows valley-side slopes to locally maintain a high angle ; the mean angle of repose of the dry sediment being determined experimentally at 46°. The comparatively low permeability promotes quite a high proportion of run-off, and the rapidity of erosion is everywhere apparent.

Records for the weather station at Hartford, 12 miles south of the area, show a mean annual precipitation of 43·09 inches over a period of 70 years, with a mean monthly maximum precipitation in 24 hours of 4·44 inches (U.S. Weather Bureau, 1952). The mean annual temperature, calculated over a 26-year period, is 49·9° F., with an absolute maximum range of from — 18° F. to 101° F. (U.S. Weather Bureau, 1930) ; the average frost season lasting from 14th October to 20th April.

Lighthouse Hollow (Text-fig. 1) exhibits a dendritic drainage system which is developing in the homogenous, easily-eroded terrace sediment. At the heads of the first order tributaries amphitheatre-like " hoppers " (Strahler, 1950, p. 803) sharply intersect the flat terrace surface, and V-shaped gulleys descend from these to the progressively deeper valleys of the higher order streams. The highest orders of streams have generally developed miniature flood plains. In all instances, however, the steep valley-sides are quite well differentiated where they intersect both the terrace summit and the flood plains. The highest order trunk segment was not investigated beyond the terrace margin, from the steep edge of which it meanders about half a mile to its junction with the Farmington River.

This network of ephemeral drainage was mapped by the first author with chain and Brunton compass, with measurements being taken at intervals of up to 70 feet, but averaging closer to 10 feet along the lower order stream segments. In addition, 1,207 complete valley-side slope profiles were measured adjacent to stream segments of all orders ; giving 750 first order side slopes, 336 second order, 85 third order, 21 fourth order, 10 fifth order, and 5 sixth order. These measurements were read to the nearest ½ foot with the chain and slope angles were obtained to the nearest degree. On short or irregular slope segments a 2-foot board was placed on the slope to mask the micro-relief. The surveys of valley-side slopes were made along true orthogonals from the basal stream channel or flood plain edge to the well-defined divide or terrace surface, avoiding spurs and those slopes on which incipient rill channels seemed to be developing.

Streams were designated by order according to the method proposed by Horton (1945, p. 281), as modified by Strahler (1952, p. 1120), whereby fingertip channels without tributaries are classed as first order streams and the junction of two streams of the same order begins the segment of the next higher order. Lighthouse Hollow is thus a sixth order drainage basin, and included within a total basin area of 1,354,800

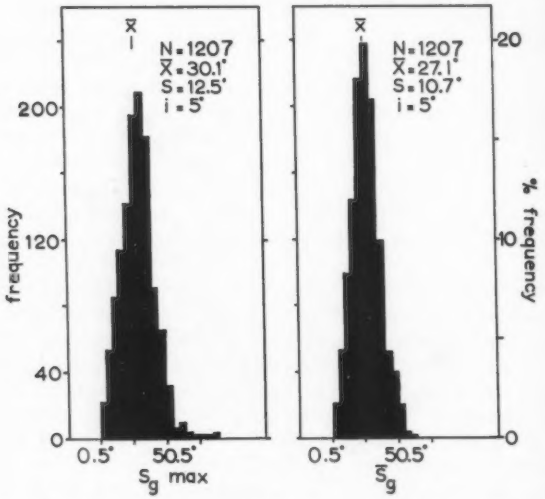


TEXT-FIG. 1.—Lighthouse Hollow drainage network. Generalized contours show the terrace form.

square feet are 29,315 feet of drainage channels. This gives a drainage density of 114 (miles per square mile) for the basin as a whole, compared with 77·6 and 258 for the Chadron and Brule formations, South Dakota, respectively (Smith, 1958, p. 1001), and 602 for the Perth Amboy badlands (Schumm, 1956, p. 612). The difficulty of accurately outlining the watershed of Lighthouse Hollow on the gently-sloping terrace surface may have led here to an underestimation of its effective drainage density. The "Constant of Channel Maintenance" for Lighthouse Hollow thus shows a requirement of 46 square feet to support 1 foot of drainage channel, and this compares with a value

of 9 for Perth Amboy (Schumm, 1956, p. 607). The small scale of the topography is emphasized further by the low mean length of first order streams of 22·77 feet, which may be compared with the 10·1 feet mean length of first order Perth Amboy streams.

Valley-side slopes (S_g) were treated statistically in two ways. Firstly, by employing the maximum angles of slope (S_g max.) along each profile



TEXT-FIG. 2.—Frequency distribution of maximum and mean valley-side slope angles in Lighthouse Hollow.

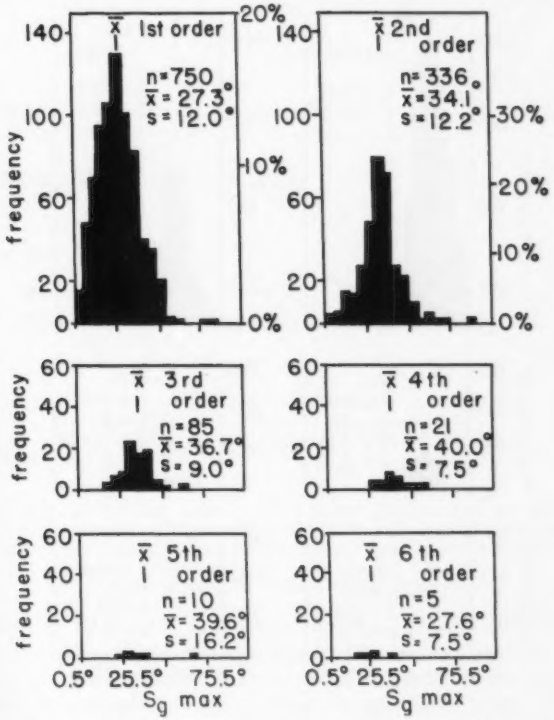
(Strahler, 1950) and, secondly, by the use of the mean valley-side slope angle ($\bar{S}_g$) of readings taken along each orthogonal from the channel or floodplain edge to the terrace surface. Text-fig. 2 shows the characteristically normal frequency distribution for these maximum and mean valley-side slope angles, the latter forming an adjunct to Strahler's Law of Constancy of Slopes:—

> " Within an area of essentially uniform lithology, soils, vegetation, climate and stage of development, maximum slope angles tend to be normally distributed with low dispersion about a mean value determined by the combined factors of drainage density, relief and slope-profile curvature ". (Strahler, 1950, p. 685.)

A " t " test (Croxton and Cowden, 1939) naturally showed a significant difference between the means of the above distributions, at the critical level of significance P = ·05, but the large proportion of short first

order valley-side slopes, where mean and maximum slope angles are not as dissimilar as elsewhere, explains the small difference between the means shown in Text-fig. 2.

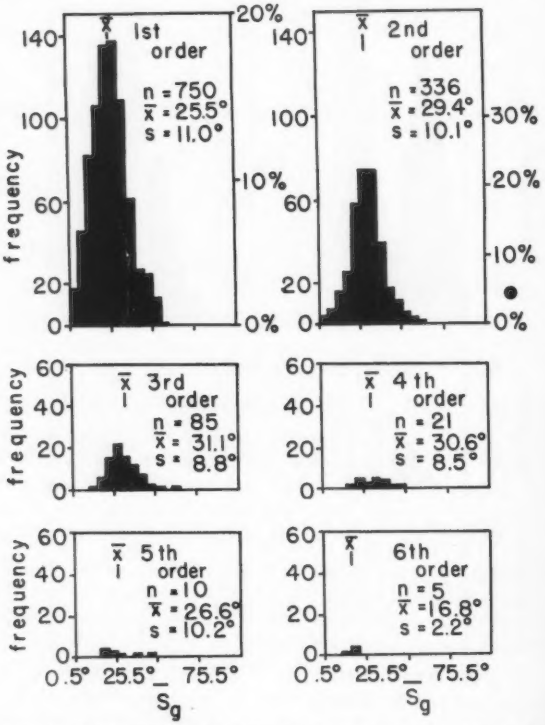
The drainage system is an actively expanding one, not bounded on its immediate perimeter by other systems, and limited only by its own erosive ability to date. Most of the terrace, into which it is incising itself, has the pristine appearance of the ideal initial surface often assumed for the Davisian " cycle ". After a heavy rainstorm the upper tributary heads are often choked with washed and slumped terrace sediment and even fallen trees, and gulleys can thus be observed to constantly invade new territory, ever extending the boundaries of the basin. This progressive cutting back of the system is perhaps significant because it implies a change in the absolute basin morphometry with



TEXT-FIG. 3.—Frequency distributions of maximum valley-side slope angles by order.

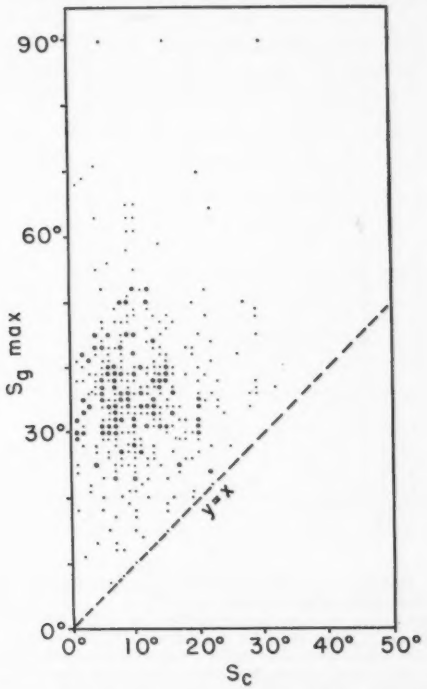
time, although investigation has shown here that the ratios which form the bases of what have been termed the "Laws of Morphometry" (Schumm, 1956; Chorley, 1957) apply just as strikingly to this expanding stream system as to one which has achieved maximum extension. However, it does seem that here the time element can be introduced into the study of drainage basin development, on the assumption that certain elements of the system are younger in stage of development than others, and that progressively varying time sequences are represented among them. It is in an attempt to determine if, and in what manner, valley-side slope angles change that such time stages are here inferred.

Maximum and mean valley-side slope angles have been grouped according to the order of the basal stream segment (Text-figs. 3 and 4).



TEXT-FIG. 4.—Frequency distributions of mean valley-side slope angles by order.

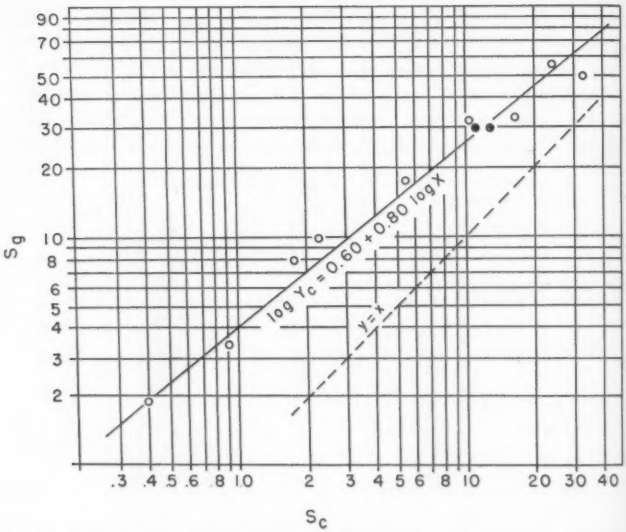
The employment of absolute frequencies serves to indicate the unavoidably disproportionate number of variates within each order, due to the frequent breaching of the comparatively short stretches of higher order valley-sides by the entrance of many tributaries, and to the increasing frequency of incipient rill development on the longer



TEXT-FIG. 5.—Relation between 433 channel gradients and the associated maximum valley-side slope angles for all orders except the first. (twenty-four suspect readings lying below the limiting ratio $y = x$ have been omitted).
Large dots indicate more than one coincident reading.

slopes of higher order. In addition, the ordering of the channel network was not carried out until the channel and slope survey was completed, and it was only then apparent that the randomly made valley-side slope measurements showed a deficit for the higher orders. Normal distributions, already noted for all slope angles, seem to be maintained within groups for each order, although a tendency for right-skewness

in the maximum slopes of lower orders reflects recent slump scars in the zone of present active extensions of the system. The small size and scoop-shaped nature of the first order valleys often presented a problem of identifying truly orthogonal valley-side slopes, and even, in many instances, of distinguishing between the valley-side slope and the stream channel. For these reasons the low maximum and mean first order



TEXT-FIG. 6.—Regional plot of mean valley-side slope against mean channel gradient (after Strahler, 1950, p. 689).
Relationship between mean valley-side slope angle and weighted mean channel gradient for Lighthouse Hollow shown by the solid dot for orders 2–6, and by the crossed dot for second order slopes. The open dots represent Strahler's regional data, on the basis of which the regression line was calculated by him.

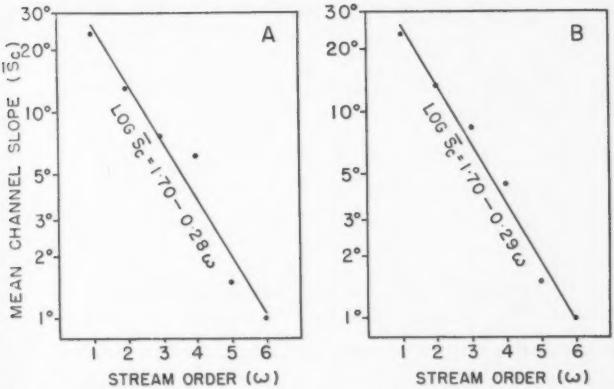
valley-side slope angles should not go unquestioned, and in the following treatment it has been considered wise to omit consideration of the relationship between first order channel and ground slopes where pairs of individual measurements are involved.

VALLEY-SIDE SLOPES AND STREAM CHANNEL GRADIENTS

It has long been an axiom of Davisian geomorphology that both stream channel and valley-side gradients progressively decline with time. More recent workers have been impressed by the possibility of a relationship existing between the gradient of the basal stream channel

(S_c) and the slope of the adjacent valley-side—the former possibly being the independent variable. Horton's " slope ratio " (1945, p. 285) $\left(\frac{S_c}{S_g}\right)$ expresses this relationship, which Strahler (1950, p. 688) stated as :—

" Ground and channel gradients . . . achieve a form best adapted to maintaining a steady state in the removal of debris . . . one would expect the angle of slope of valley walls to vary systematically with gradient of channel at the slope base. Steep ground slopes would be expected to correspond with steep channel gradients ; low ground slopes with low stream channel gradients ".



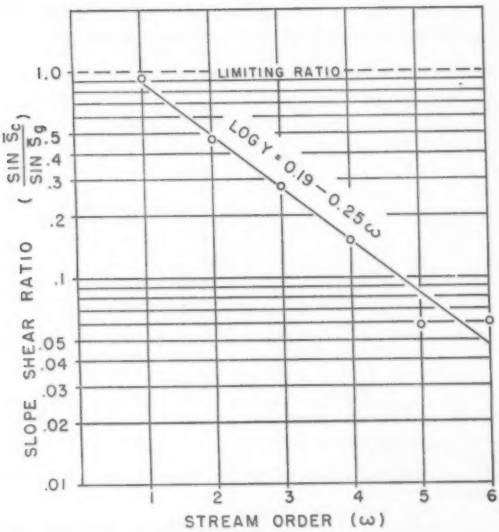
TEXT-FIG. 7.—Plot of mean channel slope versus stream order for Lighthouse Hollow.
A. From 1207 point samples.
B. From completely surveyed profiles of 610 first, 189 second, 39 third, 7 fourth, 2 fifth and 1 sixth order streams.

A plot of individual maximum valley-side slopes versus associated basal channel slopes (Text-fig. 5) certainly shows no such straight forward relationship for Lighthouse Hollow. When, however, weighted mean values of mean valley-side and stream channel gradients for the area as a whole are plotted on Strahler's graph (1950, p. 689) (Text-fig. 6) they are found to conform strikingly to the generalized relationship suggested by that author from similarly averaged figures for the second order gradients of nine other regions. Thus the stream channel and valley-side " slope ratio " concept seems to have validity in a regionally statistical sense, rather than when applied to all individual instances.

Thus the angles of valley-side slopes are controlled, at least partly, by the gradients of the stream channels at their base (Strahler, 1950,

p. 687). Removal of debris on a slope or in a channel is proportional to the downslope gravitational force, which can be expressed by the sine of the slope angle (Schumm, 1956, p. 627):—

" Thus the sine of angle of ground slope represents that part of the total gravitational force which tends to produce downhill sliding or flowage of rock particles or fluids on the surface ". (Strahler, 1956, p. 577.)



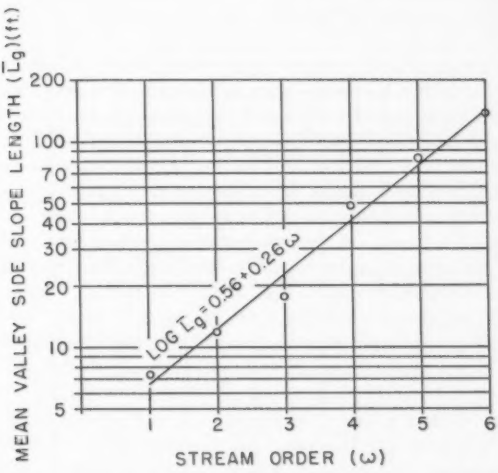
TEXT-FIG. 8.—Sine slope shear ratio versus stream order.

Employing mean channel slopes (Text-fig. 7b) and mean valley-side slopes (Text-fig. 4) a sine slope ratio was obtained for each order, similar to Horton's slope ratio. This ratio, which is a measure of relative downslope gravitational shear stresses in the channel and on the valley-side slope, bears a strikingly negative exponential relationship to stream order (Text-fig. 8). A variation of the slope ratio with the magnitude of the angles involved was noted by Strahler (1950, p. 688):

" . . . ground slope varies as somewhat less than the first power of channel slope . . . the slope ratio tends to range from 1/5 to 1/4 for low values of slopes, up to about 1/2 in the higher values ".

The limiting upper value of unity (Strahler, 1950, p. 689), which also applies to the sine slope shear ratio, is naturally most nearly approached at the heads of the first order hoppers where ground slope and channel

slope merge with one another. Because the ratio of the sines of the slopes is not constant with changing order, one may, perhaps, be allowed to speculate on the nature of the additional factor(s), itself increasing exponentially with order, which must maintain the equality of debris removal between mean ground slopes and the mean slope of the associated channel. Leopold and Miller (1956, p. 23) have shown that discharge increases in just such an exponential manner with order.



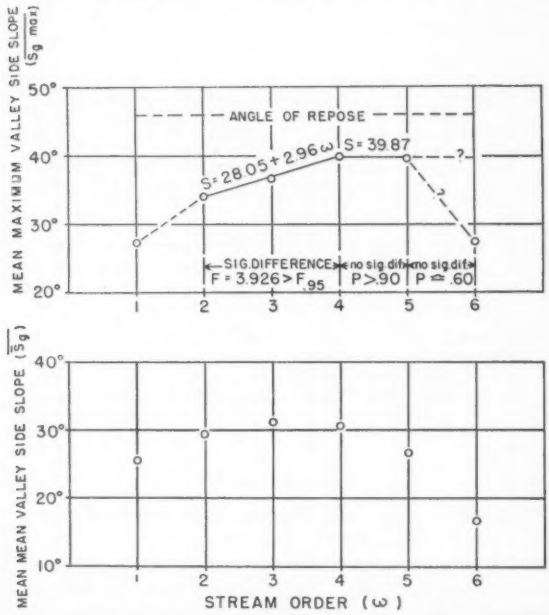
TEXT-FIG. 9.—Mean valley-side slope length versus stream order.

It is possible that this increase of available discharge in a stream channel as the headward working of the tributaries both increases the contributing basin area (Leopold and Miller, 1956, p. 20) and, by branching, converts it into a segment of higher order (Schumm, 1956, p. 621), may allow for an increase in channel removal potential so that channel and slope debris movement are in harmony for all orders. Adjacent valley-side slope angles thus, under given conditions, may be sympathetically increased with order, as they are encouraged to supply more material directly to a stream segment into which a progressively increasing discharge is being channelled. Even the consequent lowering of the slope of the channel by continued downcutting would do little to counteract the ability of water flowing in the channel to increase with order the removal rate of material from the adjacent valley-side slopes, for it has been shown (Leopold, 1953, p. 610–611) that the velocity of stream flow is controlled to a very far greater extent by flow depth than

by channel slope, and that depth of flow is a direct function of discharge (Leopold and Maddock, 1953).

The Development of Valley-Side Slopes

The drainage network of Lighthouse Hollow is cutting back into the terrace by the headward extension of the first order tributaries and by the formation of new fingertip channels where downward incision forms



TEXT-FIG. 10.—Mean maximum and mean valley-side slope angles versus stream order.

an adjacent valley-side slope (Text-fig. 9) (" Length of Overland Flow ") which is longer than the appropriate " belt of no (sheet) erosion " (Horton, 1945, p. 339), or such that the slope debris may become unstable when saturated and the slump scar encourages a new gully to develop. During the development of such a system, a stream segment of one order may be expected to be constantly and progressively changed into one of higher orders as channel extension and ramification occurs to the headward of it. It seems that it may be possible here, therefore, to employ increasing stream order as a measure of the passage of time, so making this classification of stream segments in a sense

genetic. An obvious advantage in using order in this chronological manner is that it gives an opportunity for many variates to be so grouped together and handled statistically (Strahler, 1954) so that general characteristics and trends may emerge which are not immediately apparent from individual values.

Following this reasoning, it may be considered possible, by examining the valley-side slope characteristics as a function of basal stream order, to obtain some indication of the changes in angle undergone progressively by a single slope as it finds itself successively yoked to a channel of constantly increasing order. Text-fig. 10 shows mean maximum and mean valley-side slope angles plotted against stream order. Maximum slope angles progressively increase with order for the first four orders, and an analysis of variance (Walker and Lev, 1953) indicates a significant difference of means between the second, third, and fourth orders, at the critical level $F_{.95}$. This increasing trend of the lower order maximum valley-side slope angles is mirrored by the mean valley-side slopes, and perhaps can be attributed to the increase of discharge with order previously mentioned. It is interesting that Smith (1958, p. 1005) tentatively noted an increase in slope angle with slope length for short slopes of less than 45 feet marking the dissected edge of the Orella member of the lower Brule formation flanking pediment surfaces in South Dakota. It would seem unwise, however, to press this relationship because both slope angle and length would seem to be variables dependent for their magnitude upon other causal factors. Obviously valley-side slope angles cannot continue to increase with order indefinitely, being limited by the angle of repose of the slope material in its natural state of moisture, cohesion, and vegetative binding. Schumm (1956, p. 635) has suggested that parallel retreat of slopes may occur, after a regionally characteristic (repose ?) slope angle has been attained, as long as continued basal stream downcutting or debris removal takes place. A " t " test (Croxton and Cowden, 1939) indicates no significant difference between the means of maximum valley-side slope angles for fourth and fifth orders, and this parallelism is also apparent between third and fourth order mean valley-side slope angles. Valley-side slope trends in the higher orders of Lighthouse Hollow are statistically obscured by the small sizes of the samples, and, despite the apparent low values of sixth order slope angles, a " t " test involving small samples indicates no significant difference between the means for maximum fifth and sixth order slope angles. From his work in the Verdugo Hills, Strahler (1950, p. 813) has concluded, however, " . . . that the slopes which have been protected from recent basal cutting have significantly lower angles " and thus one might be tempted to associate the apparent decline in angle of valley-side slope of the higher orders with progressive slope reduction by creep and

sheetwash after the initiation of lateral stream channel shifting at the slope base has allowed deposition there.

Steepening, parallel development, and reclining retreat of a given valley-side slope apparently all have their place in such a scheme of erosional development characterized by headward channel extension and ramification.

ACKNOWLEDGMENTS

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Some supposed 'raised beach' deposits at Corran, Loch Linnhe, and Loch Etive

By S. B. McCann

Abstract

Two large gravel deposits, at Corran on Loch Linnhe, Inverness-shire, and at the entrance to Loch Etive, Argyll, are described, and it is considered that they are fluvio-glacial fans, marking a contemporaneous halt in the retreat of the valley glaciers which occupied the sites of these two lochs. Orientation analysis of the pebbles in the Corran deposit supports this conclusion. Similar analysis on gravels along the northern shore of Loch Etive indicates that they are not raised beaches, but were deposited in lakes held up by the glacier which deposited the Loch Etive fan.

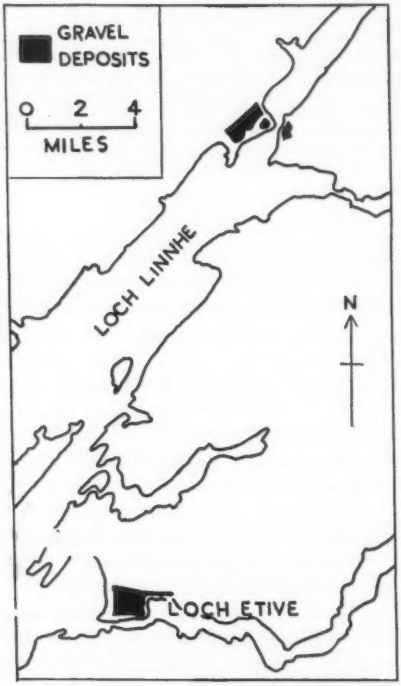
I Introduction

The association of the higher raised beaches of western Scotland with the last stages of glaciation has long been recognized, and W. B. Wright (Wright, 1937) concluded that the late-glacial period of high sea-level had fallen from its maximum before the final retreat of the valley glaciers. In his account of the evidence for this conclusion, Wright describes the great deposit of gravels at Corran Ferry, which form a partial barrier across Loch Linnhe (Text-fig. 1). Earlier, E. B. Bailey (Bailey, 1916) had noted the absence of the 100 foot beach in this portion of Loch Linnhe and recognized that the Corran gravels, lying at about 75 feet above sea level, are quite different from other fluvio-glacial features in the area. He accepted the view put forward by another survey officer, Grant Wilson, that the gravels mark the site of a lake held up between two glaciers, one from Loch Linnhe, the other from Loch Leven. This theory was accepted by Maufe, and Peach and Horne, the last authors suggesting (Peach and Horne, 1916) that the lakes on the Corran peninsula mark the sites of buried ice, which persisted during deposition. Wright took a similar view that the gravels are not a raised beach spit, but the remnants of the outwash gravel of a glacier.

The present paper consists of a re-examination of the Corran gravels and their mode of deposition, and also of a similar deposit at the entrance to Loch Etive, Argyll, which again forms a partial barrier across the loch. Descriptions of raised beaches in the Loch Etive area in the *Memoir* for Oban and Dalmally (Kynaston and others, 1908) are misleading, for the term 50-foot raised beach is used in reference to what is generally known as the 25-foot beach. The Loch Etive deposit is described as the sands and gravels of this beach, despite the fact that at its highest it is over 80 feet above the sea.

II The Corran Gravels

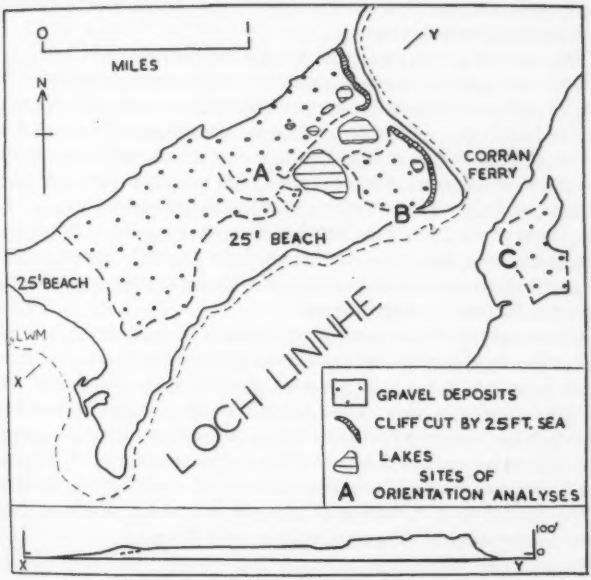
The extent of the gravel deposits at Corran is indicated on the 1-inch Geological Map, Sheet 53. The deposits form the major part of the peninsula into Loch Linnhe and are bordered by the deposits and eroded terrace of the later 25-foot sea. No mention is made in the



Text-fig. 1.—Location map.

Memoir description of the precise height of the gravels, nor of the form of the deposit. Measurement, using an Abney hand level, shows that the surface of the deposit on the north eastern part is 77 feet above the high water mark of ordinary spring tides (H.W.M.), and that there is a gradual decline in height to 40 feet in the south western portion. (Text-fig. 2 and section). The hollow formed by the two large lochans divides the gravel mass into two parts, but this does not destroy the unity of form shown by the deposit, namely that gradual slope of the

whole outwards down Loch Linnhe. A shallow valley has been cut into the main part of the deposit by a stream draining the steep mountain slopes to the west, but the main modifications on the level sloping surface of the deposit are the circular hollows now occupied by lakes. The largest of these are 74 feet and 43 feet deep, and the Ordnance Survey gives the surface levels as 14 feet and 6 feet respectively (Bailey, 1916), but the other four are much smaller and merely pit the surface of the deposit.



TEXT-FIG. 2.—The Corran gravels, Loch Linnhe.

The relations of the peripheral 25-foot beach to the older gravels is a direct result of the outward sloping nature of the main deposit. Along the inner north-east facing shore of the peninsula the 25-foot sea has cut a narrow, well-marked terrace into the deposit where it is at its highest. The angle formed between this terrace and the backing cliff is 35 to 38 feet above H.W.M. On the seaward south-western part of the peninsula, on the other hand, the 25-foot beach is depositional, the highest beach ridge of this period being traceable beneath the peat at about 40 feet above H.W.M., overlying the lower south western part

of the older gravels (Text-fig. 2, section). Similar 25-foot beach deposits underlie the fields at each end of the vale formed by the two large lochans. From a maximum height of 38 feet, the 25-foot sea retreated to the present sea level and it is possible to trace various stages in this retreat from the form of the old beach ridges, which can be seen in the fields of Clovullin crofts.,

The form of the gravel deposit and its later modification by the 25-foot sea is shown also by the submarine contours. On the inner side of the peninsula there is a steep submarine slope to the floor of the loch, while on the seaward south-western side there is a wide foreshore and gentle submarine slope.

The Corran peninsula is thus wholly depositional, consisting of a deposit of gravels sloping outwards and south-westwards down Loch Linnhe, and is modified peripherally by the erosion and deposition of the 25-foot sea. A similar, though smaller, deposit with a uniform gently sloping surface at about 66 feet on the east side of the loch at Corran Ferry suggests that the deposit once formed a complete barrier across the loch. The breaching of this barrier occurred before the maximum of the 25-foot sea, for deposits and erosional terraces of this sea occur along the inner portion of Loch Linnhe. This breaching was probably aided by the presence of a through valley similar to that occupied by the two large lochans.

It is suggested, on the basis of the physical form of the deposit, that it is a fan deposited by streams from a glacier, which extended down Loch Linnhe and had a long period of stillstand at this point. The outward sloping surface of the deposit is characteristic of outwash fans and the steep inner north east facing slope is considered to represent the original ice contact slope, modified by later marine erosion during the 25-foot sea period. The deep hollows are considered to be kettle holes, formed by the melting of large masses of ice which persisted during deposition, as suggested by Peach and Horne.

Several large exposures occur on the Corran peninsula and the deposits consist of ill-sorted sands and gravels with indications of very rapid deposition. It is not always possible to see clearly defined stratification within the deposits and bedding is very variable, with dips of 25 degrees to 30 degrees to the west. Bedding planes are well defined for short distances and then all traces disappear. All sizes of roughly rounded pebbles and small cobbles occur within a small area in a matrix of coarse sand, which in places includes a considerable proportion of fine material. This sand and mixed silt also occurs as lenses, which sometimes show current bedding and fine stratification. An analysis of the orientation of the pebbles within the deposit, described below, provides additional evidence for the proposed fluvio-glacial fan origin.

Orientation Analysis

W. C. Krumbein (1939) has described methods of measuring and analysing the preferred orientation of pebbles in a deposit. The most common method is to use the dip direction and angle of dip of the longest axis of the pebbles, but for a more complete statement Krumbein suggested the use of the longest axis and the maximum projection plane. However, in his analysis of the flood gravels of San Gabriel Canyon and Arroyo Seco, California, he used only the data for the long axes, which he considers sufficient under ordinary circumstances (Krumbein 1940 and 1942). The data from San Gabriel Canyon and Arroyo Seco show that the long axes of stream pebbles lie parallel to the valley trend and that the majority dip upstream :—

"Pebble orientation showed uniform upstream imbrication even in the steep upper part of the canyon" (Krumbein 1942, p. 1401). The long axis is considered to be oriented parallel to the axis of movement, and to mark the direction of the current at the point of deposition. Pettijohn, after reviewing the results of various workers in this field, concluded that the longer axes of stream pebbles usually lie in the direction of the current (Pettijohn 1957).

Measurements of pebble orientation were made at Corran, samples being taken at sites A and B in the main deposit, and at site C on the eastern side of the loch (Text-fig. 2). The results are shown in the conventional manner in Text-fig. 3. Dip direction and angle of dip of the long axes of individual pebbles are plotted on a polar co-ordinate grid, and the two properties are shown separately on histograms. The dip direction histogram is in the form of a "rose" diagram, with the dip directions expressed as a percentage frequency within twelve classes of 30 degree interval.

The samples from Corran show a fabric which is characteristic of stream deposits, namely the general alignment of the long axes of the pebbles in one dominant direction, which has been shown by Krumbein to mark the direction of the depositing current. This direction is probably indicated best by the mean aximuth. In order to calculate the mean azimuth it is necessary to arrange the data so that classes on opposite sides of the circle are combined (Table I), for in terms of azimuth alone such classes are equivalent. The mean azimuth at site A is 94°, at site B is 89° and at site C is 95°. Thus in terms of aximuth alone the orientation of the pebbles is in accordance with the proposed origin of deposition by streams flowing in a general westerly direction from the Loch Linnhe glacier. However, within the general east-west orientation of the long axes, more pebbles dip to the west than to the east, so that if the rule of upstream imbrication is applied it would appear that the deposits were laid down by streams flowing in the opposite direction. The evidence of stratification at these sites precludes such a possibility,

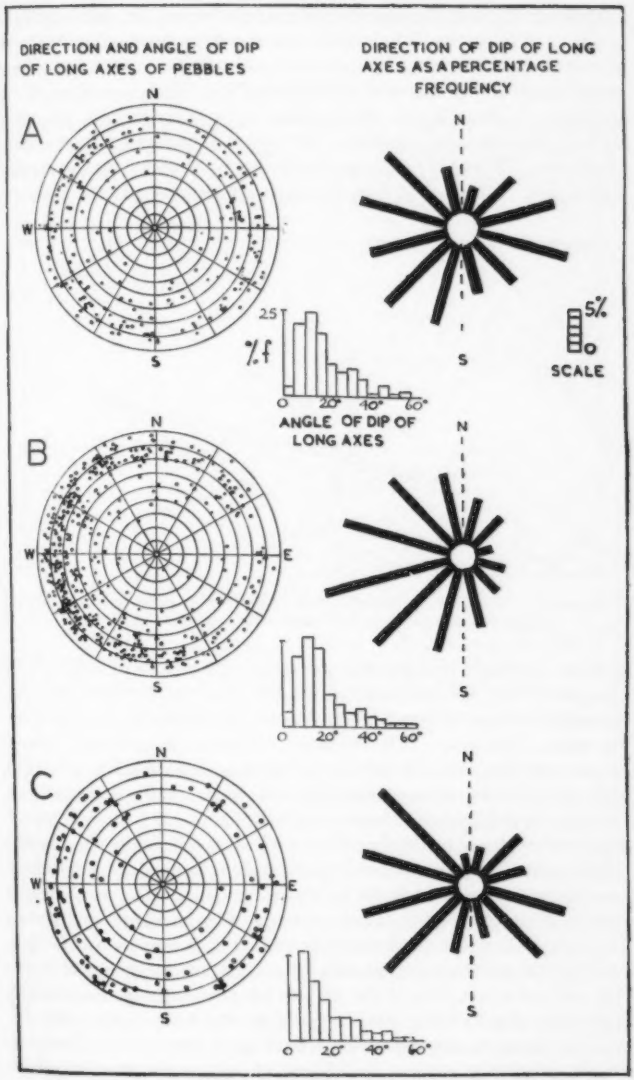
for where traces of bedding can be seen the dip is to the west at angles of 20 to 25°. The deposition of the gravels by west-flowing streams is thus confirmed, and the high dip of the beds, which is an original sedimentary feature provides an explanation of the downstream imbrication.

The dip of the long axes of an individual pebble measured in the field, and indicated on Text-fig. 3, is the absolute dip in relation to a horizontal plane. If, however, this dip is considered in relation to the

TABLE 1.—DIRECTION OF DIP OF LONG AXES OF PEBBLES, EXPRESSED AS A PERCENTAGE FREQUENCY WITHIN SIX COMBINED CLASSES OF 30° INTERVAL.

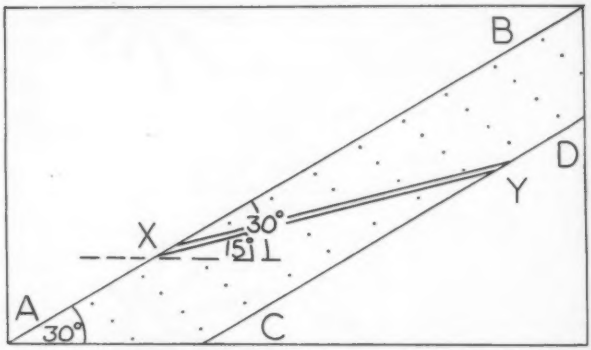
Combined Classes	Percentage frequency in each class					
	Corran			Loch Etive		
	A	B	C	R1	R2	S
(1–30°) + (180–210°)	13	15·5	10	11	13	16·66
(31–60°) + (211–240°)	17·5	18·25	20	12	12	10·66
(61–90°) + (241–270°)	18·5	17·75	17	23	13	19·33
(91–120°) + (271–300°)	21·5	17	22	22	25	20·66
(121–150°) + (301–330°)	17·5	15·5	24	20	23	14·66
(151–180°) + (331–360°)	11·5	16	7	12	14	18

bed in which it lies and which dips in the same direction but at a greater angle, then the pebble can be said to dip upstream within its environment of deposition. In Text-fig. 4, AB and CD are the top and bottom of a bed dipping at 30° from right to left, which is also the direction of the depositing stream. XY represents the long axis of a pebble in this bed which dips in the same direction, but at only 15°. The stream in depositing bed ABCD flowed down an original 30° slope upon the earlier deposits below CD. In relation to the plane of this bed, the pebble axis XY dips upstream 15° (angle XYC = 15°). The upstream imbrication of stream pebbles represents the position of least resistance to the current, and therefore of greatest stability. Under conditions with a depositional gradient of 30°, a dip of a pebble at 15° in the same direction represents somewhat analogous conditions of



TEXT-FIG. 3.—Orientation analysis data from the Corran gravels.

stability to a dip in the upstream direction of 15 to 20°, with a stream gradient of 5° or less. The apparent downstream imbriçation becomes a true upstream imbrication if considered in relation to the immediate environment of deposition of an individual bed. The orientation of the pebbles at the Corran sites where 65 per cent, 68 per cent and 70 per cent, dip at less than 20° within beds dipping in the same direction at 20 to 25° suggests conditions of stability during stream deposition. Wadell described a similar orientation pattern with the



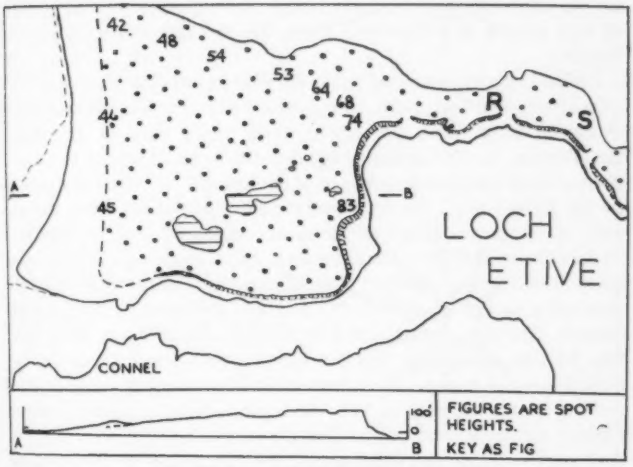
TEXT-FIG. 4.—Diagram to illustrate the attitude of a pebble within bedded gravels deposited by a stream with a gradient of 30°.

pebbles dipping in the direction of the bed, in a section of esker gravel (Wadell 1936). He concluded, by analogy with talus slopes, that the gravel was a subaerial gravitational deposit, which had rolled down from the crest of the esker. The evidence at the Corran sampling sites of current-bedding and fine stratification precludes such an origin. The gravels were definitely deposited by flowing water and the fabric has been shown to be compatible with stream deposition. It is considered that the gravels were deposited by fast-flowing streams with irregular channels, which issued from the snout of the Loch Linnhe glacier. The streams carried great amounts of debris, which was deposited with little sorting, as a series of steeply dipping discontinuous beds in the rear part of the fan. It is these beds which can be examined at the sites shown on Text-fig. 2, and from which the measurements for orientation analysis were taken. The marked orientation of the pebbles within the gravels excludes the possibility that the fan was deposited in the Late Glacial sea, for marine transport and deposition would have destroyed this primary stream orientation. It can be inferred, therefore, that the late glacial sea level had fallen from its maximum, during which the 100-foot

raised beaches of western Scotland were formed, before the deposition of the Corran gravels.

III. The Loch Etive Gravels

The peninsula at the entrance to Loch Etive consists essentially of a great deposit of gravels, the surface of which slopes outwards, down the loch, from 83 feet above H.W.M. in the east to 40 feet in the west (Text-fig. 5). The 25-foot raised beach forms the margins of the



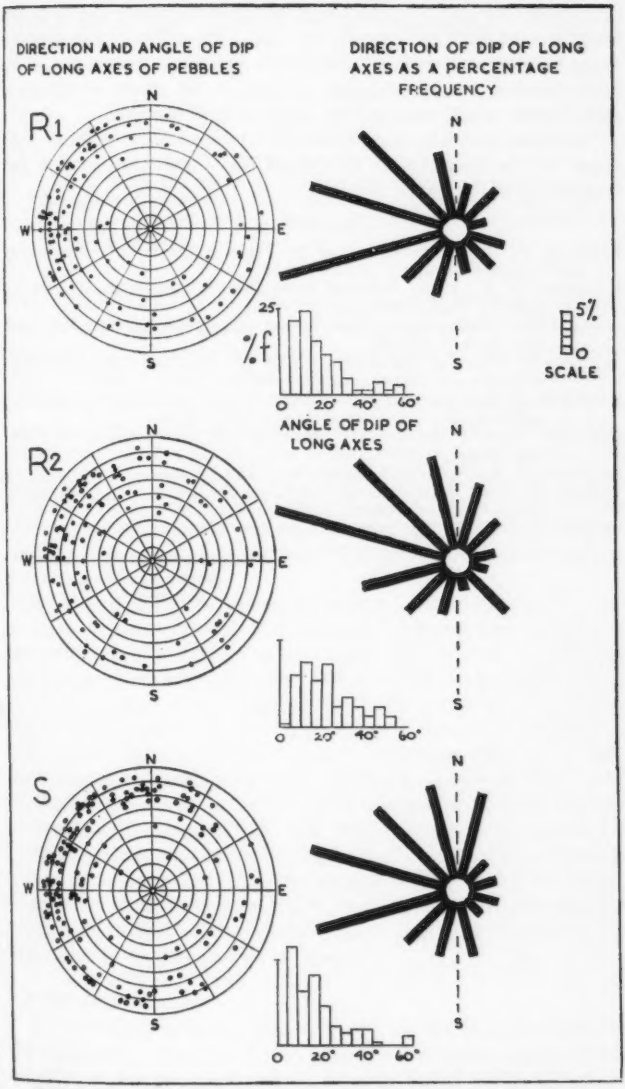
Text-fig. 5.—The Loch Etive gravels.

peninsula. On the inner eastern and southern shores the 25-foot sea has cut a narrow terrace, backed by a steep cliff, into the higher part of the older gravels, but on the outer western shore the 25-foot beach is depositional. The highest beach deposits of this period can be seen, overlying the older gravels, at several places near the main Oban-Fort William road. Cuttings along the railway, near the western shore of the peninsula, are in the higher beach deposits of the 25-foot sea, which reach 45 feet above H.W.M. As an eroded feature, the maximum height of the 25-foot beach, at the angle between the old cliff and the old shore platform, is about 35 feet.

A thick cover of peat overlies a large area of the older gravels, but the uniform west sloping surface is broken by two large and several smaller hollows, containing lakes similar to those at Corran. The analogy of the hollows in the Loch Etive and Corran peninsula is

paralleled in the general form of the deposits, and in the later modification by the 25-foot sea. No exposures suitable for orientation analysis occur within the main fan deposit at Loch Etive, but the evidence from numerous small exposures, especially in the cliff cut by the 25-foot sea, shows that the gravels are similar to those at Corran. It is suggested, therefore, that the Loch Etive deposit is also a glacial fan, marking a halt in the retreat of the Loch Etive glacier, contemporaneous with that of the Loch Linnhe glacier at Corran. In one detail the Loch Etive deposit differs from that at Corran, for the exposure of solid rock on the north side of the railway bridge at Connel and the occurrence of rock islands in the narrows show that the gravels rest on a rock barrier.

Gravel deposits, described in the *Memoir* as 100-foot beach deposits (Kynaston and others, 1908), form a narrow platform along the northern shore of Loch Etive for about three miles east of the main fan (Text-fig. 5). The surface of the deposits varies in height, but has a general slope southward to the top of the low cliff, cut into the deposits by the 25-foot sea. The height of the platform, at the inner margin with the mountain slopes to the north, varies between 75 feet and 95 feet above H.W.M. Measurements for orientation analysis were made at two locations within these deposits ; at site R measurements were made in two gravel pits (R1 and R2) and at site S in a stream section. The dip of the bedding at R is variable and individual beds may become attenuated, but the dip is always in a south-easterly direction at low angles. Measurements in both pits were made in beds dipping in this direction at 11°. At S the bedding is fairly constant, dipping east-south-east at 8°. The dip of the bedding is an original sedimentary feature and indicates that the direction of transport of the gravels during deposition was from north-west to south-east. The data for the direction and angle of dip of the long axes of the pebbles, from both sites, show a marked preferred orientation, which is characteristic of stream deposits (Text-fig. 6); and the mean azimuths indicating the direction of the current at the time of deposition are R1–101°, R2–98° and S—93° (calculated from the combined azimuthal classes shown on Table 1). Within this pronounced orientation pattern, the majority of the pebbles dip in the opposite direction to the dip of the beds, exhibiting the normal upstream imbrication described by Krumbein. Thus the fabric of the gravels along the northern shore of Loch Etive provides strong evidence that they are not marine gravels, but were deposited by streams flowing from the north-west. Nor is it likely that this deposition took place in the 100-foot sea, for if this had been the case, then the primary orientation of the pebbles, due to the stream action, would have been considerably modified by the reworking of the gravel by the sea. It is suggested, therefore, that the gravels were deposited in a lake held



TEXT-FIG. 6.—Orientation analysis data from the Loch Etive gravels.

up between the glacier to the south, the steep mountain slopes to the north and the Loch Etive fan to the west. Thus all the gravel deposits along the northern shore of Loch Etive, previously considered to be raised beaches, can be related to the halt in the retreat of the great valley glacier which occupied the present site of the loch.

The author wishes to acknowledge the advice given by Professor J. A. Steers of the Department of Geography, Cambridge, during the preparation of this paper.

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Lower Devonian Crinoids from Toongabbie, Victoria, Australia

By G. M. Philip

(PLATE VIII)

Abstract

Three camerate crinoids, *Thylacocrinus*(?) *ignotus* sp. nov., *Eucalyptocrinites inchoatus* sp. nov. and *Hexacrinites*(?) sp. are described from the lower Devonian of Toongabbie, Gippsland, Victoria, Australia.

Introduction

Very little is known of the Palaeozoic crinoids of Australia, particularly the camerate crinoids. The five named species of camerate crinoid described from various localities in Eastern Australia are poorly characterized specifically and probably all unidentifiable generically.

From the Upper Silurian Melbourne Beds two of these species have been described. They are *Hapalocrinus victoriæ* Bather, 1897 (p. 337, pl. 15), a generic determination questioned by Clark (1946), and *Helicocrinus plumosus* Chapman, 1903 (pp. 107–8, pl. 17; pl. 18, figs. 1–5), the type species of *Helicocrinus* Chapman, 1903 (*non* Quenstedt, 1876), a camerate crinoid of uncertain affinities. Both these species are based on single specimens preserved as moulds in which the plating is poorly defined.

The Upper Palaeozoic record is as poor. Etheridge (1892, p. 78, pl. 22, fig. 4) has described " *Periechocrinus* " *indicator* from a fragmentary specimen of probable Carboniferous age, a further questionable generic determination. Etheridge (*in* Jack and Etheridge, 1892, p. 208, pl. 38, fig. 3) again named a singularly imperfect cup from the Carboniferous of Queensland *Platycrinus* (?) *nux*. Various authors (McCoy, 1847, p. 229; de Koninck, 1877, p. 160; Etheridge, 1892, p. 77) have also recorded fragmentary remains of *Actinocrinus* collected from rocks of Carboniferous age in N.S.W. and Queensland. More recently Sieverts-Dorek (1942, p. 225, figs. 1–3) described *Camptocrinus*(?) *tasmaniensis* from the Permian of Tasmania, based on stem columnals.

It can be seen, then, that the present description of the rather fragmentary fauna from Toongabbie, Victoria, in which three camerate species are recognized, represents a quite surprising increase in the knowledge of Australian crinoid faunas.

Locality and Age.

The specimens studied were collected from the lower Toongabbie Marble Quarry, Thompson River, Gippsland, Victoria. A map of this locality has been given elsewhere (Talent and Philip, 1956). Talent has also provided a preliminary account of the crinoids (p. 74 " Pelma-

tozoa ") of the fauna of the limestone horizon in discussing its palaeoecology. The specimens available to Talent are described here, together with further material collected subsequently by the writer.

The age of this horizon has been discussed elsewhere by the writer (Philip, 1960) where it is concluded that all the major limestones of the Gippsland area in the Walhalla Synclinorium and its extensions (Tyers River, Coopers Creek, Toongabbie, Deep Creek, Loyola) are of very similar age, which, from the evidence available at present, is most probably basal Lower Devonian. The three crinoids recognized here suggest an age quite in keeping with this although they afford no precise indication of age. They are:—

1. *Thylacocrinus*(?) *ignotus* sp. nov., which is only questionably referred to that Lower Devonian genus, and in all probability represents a new rhodocrinitid genus.
2. *Eucalyptocrinites inchoatus* sp. nov., which, as far as the material permits it to be said, is a typical member of the widespread Silurian and Devonian northern hemisphere genus *Eucalyptocrinites*. Of the fifty or so species of *Eucalyptocrinites* listed by Bassler and Moodey (1943) only two were described from rocks of Devonian age. Thus the occurrence at Toongabbie suggests a Silurian rather than a Devonian age for the limestone.
3. *Hexacrinites*(?) sp. The single, fragmentary specimen can only be placed in the Devonian genus with reservation. However its occurrence in the fauna offsets the Silurian aspect given by *Eucalyptocrinites inchoatus*.

Preservation.

The specimens were collected from fairly massive crinoidal limestone, which unfortunately tended to break exposing the inside of the cup. Thus many of the specimens are internal moulds of cups in limestone, with the thecal plates lining the concave surfaces of their counterparts. In all specimens the tegmen, pinnules and usually all the free brachials are missing from the crown, attesting to a certain amount of transport and/or current action before their ultimate burial.

Plate Terminology and Symbols.

Throughout the descriptions the plate terminology and symbols used are those of Bather (1900) as slightly modified by Ubaghs (1956a, p. 518). It is considered that Wanner's (1948) and Ubaghs' (1956) criticisms of the system suggested by Moore and Laudon (1941) are sufficient grounds on which to revert to Bather's abbreviations.

Systematic Descriptions.

The specimens described here are catalogued in the Melbourne University Geology Department Type and Figured Specimen Collection

(prefix M.U.G.D.) and the Sedgwick Museum (prefix S.M.). The average measurements given in the descriptions refer to the primary figured specimen, or where this is not possible, to a specimen of comparable size.

Subclass CAMERATA
Order DIPLOBATHRIDA Moore and Laudon
Superfamily RETEOCRINICAE Ubaghs
Family RHODOCRINITIDAE Roemer

The Rhodocrinitidae, as defined by Ubaghs (1952), represent a large and diverse collection of genera ranging from Middle Ordovician to Carboniferous. In the light of Wright's (1952) re-interpretation of *Rhodocrinites* Miller, 1821, it would appear that the family could well be restricted to this and other Carboniferous genera such as *Gilbertsocrinus* Phillips, 1836 and *Cribanocrinus* Kirk, 1944. Certain of the genera listed in the family by Ubaghs, such as *Wilsonocrinus* Springer, 1926 (the type species has four infrabasals) and *Anthemocrinus* Wachsmuth and Springer, 1881 (with but one primibrachial in each ray) have no place in the family even in its widest sense. However, no satisfactory redefinition of the family would be possible without a lengthy reassessment of the generic position of many of the species at present placed in the included genera.

Genus **Thylacocrinus** Œhlert, 1878
Thylacocrinus(?) *ignotus* sp. nov.

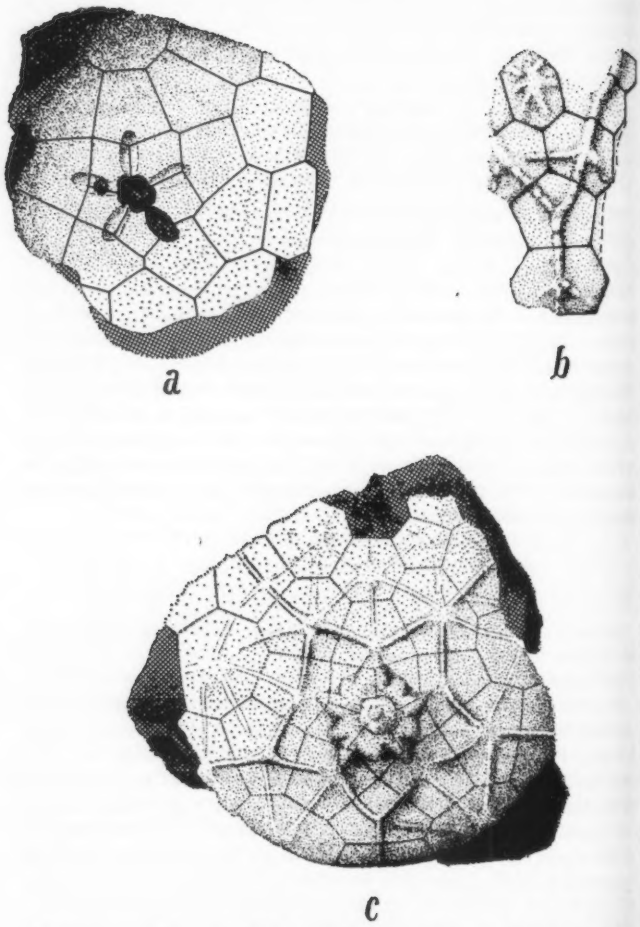
(Text-figs. 1a–c, 2, 3; Pl. VIII, figs. 1, 2, 8)

Material.—Four cup fragments (M.U.G.D. 3400–3), three of which are broken to show internal moulds only.

Holotype.—M.U.G.D. 3400.

Description.—Comparatively large, somewhat conical cup, with a rounded base, and attaining a diameter of 4 cm. at a height of about 3·5 cm. Plates up to tertibrachials incorporated in the cup. The only interruption to the basically pentamerous symmetry of the theca is in the posterior interradius, which is slightly wider than the other interradii and is made up of smaller and more numerous plates. The inner surface of the cup is marked by a well-developed system of furrows which are in general radially disposed to the plates and usually cross the sutures at right-angles. These are represented as ridges on an internal mould of the cup (Text-fig. 1c) and apparently correspond with similar external ridges on the cup surface. Of these ridges those representing ray ridges are most strongly developed.

Infrabasals.—These consist of five equal, comparatively small, basically pentagonal plates (h = 3 mm.; w = 3 mm.) grouped around a small hole at the base of the cup. Along their common sutures on the inside of the cup are five petalloid depressions which extend a short



TEXT-FIG. 1.—*Thylacocrinus*(?) *ignotus* sp. nov. a, Holotype, M.U.G.D. 3400, internal view of base of cup showing the plating. b, Fragment of brachials with iIIBr, showing the ray ridges and surface ornament of the plates, M.U.G.D. 3403. c, Internal mould of basal part of cup showing the system of radiating ridges and ray ridges with the inferred sutures superimposed, M.U.G.D. 3402. All × 2.

distance onto the basals (Text-fig. 1a). A well-preserved internal mould of the base of the cup (Text-fig. 1c) shows a prominent basal protuberance around the periphery of which are ten ill-defined ridges. The five more strongly developed of these, which alternate with a less pronounced series, correspond with the petalloid depressions mentioned above, whereas the intervening ridges probably represent the internal expression of the ray ridges on the basals. Around the periphery of the basal depression the paths of ray ridges are obscure (see also Text-fig. 3).

Basals.—These are usually pentagonal, slightly higher than wide (h = 5 mm.; w = 4·5 mm.), with greatest width along their lower edges, and with a wide, horizontal, upper suture, supporting the interradials.

Radials.—The radials are basically pentagonal (h = 5·5 mm.; w = 7 mm.), ideally with their lower apices just touching the dorsal apices of the IBB (Text-fig. 2). There is a certain amount of variation in this feature so that the IBB and the RR may be in contact for a short distance, in which case both plates develop a small extra facet, or the IBB and RR may be separated slightly by the BB, in which situation the BB develop an extra side (Text-fig. 1a). The posterior radials, where they adjoin the posterior interradius, develop a further facet on their upper edges to support the extra plates present in that interradius.

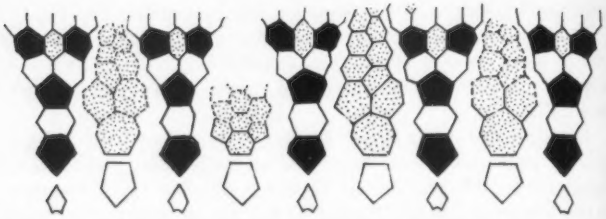
Primibrachials.—IBr_1, where known, is a fairly large, hexagonal plate somewhat wider than high (h = 5 mm.; w = 7 mm.); IBr_2 (= IAx) is a relatively higher pentagonal plate (h = 5·5 mm.; w = 6 mm.) bearing the secundibrachials.

Secundibrachials.—$IIBr_1$ is a hexagonal plate, similar in shape to IBr_1, but smaller (h = 4·5 mm.; w = 5·5 mm.), and along its adradial facet adjoins a corresponding $IIBr_1$, supported by the same IAx, and also a single iIIBr. $IIBr_2$ (= IIAx) is a smaller pentagonal plate (h = 4·5 mm.; w = 5 mm.) bearing two, probably hexagonal IIIBrr. The plating of the upper part of the cup is not shown clearly by the existing specimens, but no plates appear to be present above the IIIBrr, except, perhaps in one place on the internal impression of a cup, where there could be present an iIIIBr (shown in Text-fig. 2).

Interradials.—The interradials, particularly in the plating of their upper part, are imperfectly known, but appear to consist of a succession of 1 — 2 — 2 — 2 — 2 with the possibility of a further one or two small plates. The single large plate resting on the basals is heptagonal and is the largest cup plate (h = 7·5 mm.; w = 7 mm.). This supports horizontal series of pairs of hexagonal plates, which decrease in size distally.

Posterior Interradius.—The lower plates of this area alone are known, and then only from the internal impression of the base of a cup (Text-fig. 1c; Pl. VIII, fig. 2) which primarily shows ridges which are

the internal expression of the surface markings of the cup. However, compared with the other interradii, the posterior is slightly wider and consists of more numerous and smaller plates. The single large basal plate of the other interradii is replaced by three plates, the middle one of which alone rests on the posterior basal, with the lateral plates slightly higher. This supports a further horizontal series of three (or



TEXT-FIG. 2.—Inferred plating of *Thylacocrinus*(?) *ignotus* sp. nov. Radials and auxiliaries black; brachials, basals, and infrabasals white; interradials and interbrachials stippled; inferred sutures broken and an inferred interradial column with its supporting basal is omitted. × 1 approximately.

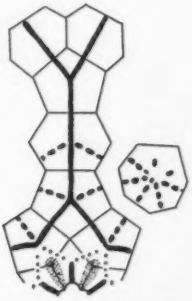
four ?) plates of similar size. Whether or not a posterior anal ridge is developed as in crinoids such as *Compsocrinus* it is impossible to say positively, but as the median ridge of the posterior interradius is not more strongly developed than the ridges of the other interradii on the inside of the cup, and certainly is much less striking than the main system of the brachial ridges, it may be inferred that in this respect the posterior interradius was not more differentiated than the others.

Interbrachials.—One fairly high, hexagonal iIIBr (h = 6 mm.; w = 4·5 mm.) is present, apparently included by the $IIBrr_1$, $IIBrr_2$ and the $IIIBrr_1$. The presence of iIIIBrr is suspected, but this cannot be established from the existing material.

Tegmen, free arms, pinnules and stem are all unknown.

Remarks.—This species can only be placed in the genus *Thylacocrinus* Œhlert with reservation. It differs from *T. vannioti* Œhlert, from the Lower Devonian of St. Germain, France, in possessing the prominent ornament of ray ridges and radiating ridges of the lower cup plates, fewer interradials and interbrachials, and apparently also lacks the peculiar posterior radii, each of which gives rise to six free arms in *T. vannioti*. In all probability *T.*(?) *ignotus* represents a new rhodocrinitid genus, but as the species is imperfectly known, and as even the description given above is based on a considerable degree of interpretation, it is best to leave the matter undecided till such time as better material is available for study.

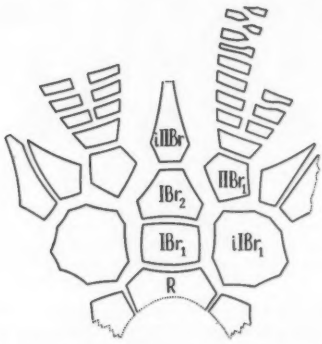
It may be noted that the North American *Thylacocrinus clarkei* Wachsmuth and Springer, 1897 (p. 248–9, pl. 13, figs. 11a, b) possesses much more complexly plated interradials and intersecundibrachials than *T. vannioti*, and has even complex intertertibrachials. It also lacks the characteristic posterior rays of the European type species. These differences are also present in *Rhodocrinus gracilis* Hall, placed by Goldring (1923) in *Thylacocrinus*.



TEXT-FIG. 3.—Schematic diagram of the arrangement of ridges on the plates of *Thylacocrinus*(?) *ignotus* sp. nov. Primary ridges represented as solid black lines, secondary ridges as broken lines, and those obscured by the basal protuberance as dots.

The pattern of radial ray ridges is of interest in *T.*(?) *ignotus*. Both Bather (1900, p. 338) and Wilson (1916, p. 549) tacitly assume in their discussions of the anal ray ridges of camerates, that the ray ridges of camerates are an expression of the nervous system. In *T.*(?) *ignotus* the ray ridges, and the primary system of grooves they give rise to on the inside of the cup, follow closely the expected paths of the brachial ectoneural nerve cords as known in living dicyclic crinoids through the work of Carpenter and Bather. Where the brachial neural system has been described in other fossil crinoid groups (in the inadunate crinoid *Calceolispongia* by Teichert, 1949, and in the articulate crinoid *Marsupites* by Sieverts, 1927) the nerve canals were found immediately beneath the inner surface of the plates, similar to their position in living crinoids. No such fine canals were seen in the plates of *T.*(?) *ignotus* when cup interiors were etched with weak acid to reveal the sutures, nor do they appear to have been reported in any other camerate. It may well be that the grooves on the insides of the plates beneath the ray ridges do in fact represent the impressions of the brachial nerve cords in camerates. A connection between plate ornament and nervous system in crinoids has been suggested by the work of Sieverts (1927) on *Marsupites*. Sieverts found that in the smoother varieties of *Marsupites*

testudinarius only the main canal system was present within the plates, whereas in the more ornamented varieties subsidiary radial canals were present paralleling the radial ornament of the plate surface. It is, however, difficult to assume that there always exists this relation between the linear plate ornament and the nervous system of crinoids; the ray ridges, and particularly the radial ornament, are repeated in many crinoid groups and even other pelmatozoans and they lack the



TEXT-FIG. 4.—*Eucalyptocrinites inchoatus* sp. nov. Plating of portion of the exterior of the holotype, M.U.G.D. 3404. × 2.

stability and persistence one would expect of a feature were it always related to such an intrinsic character as the ectoneural system.

Order MONOBATHRIDA Moore and Laudon
Superfamily DOLATOCRINICAE Ubaghs
Family EUCALYPTOCRINITIDAE Roemer
Genus **Eucalyptocrinites** Goldfuss, 1826
Eucalyptocrinites inchoatus sp. nov.

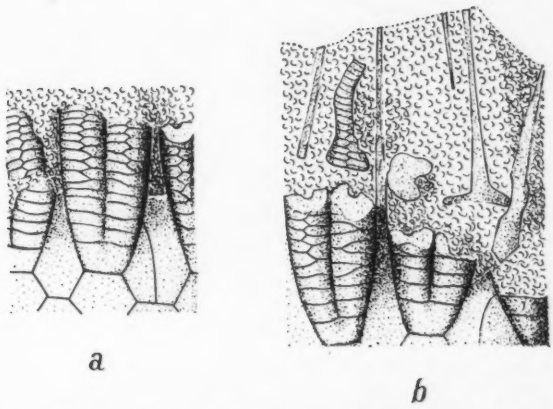
(Text-figs, 4–9; Pl. VIII, figs. 3, 4, 9)

Material.—Eight cup fragments with occasional associated columnals (M.U.G.D. 3404–10; S.M. H7524)

Holotype.—M.U.G.D. 3404.

Description.—The cup is medium sized and hemispherical with a wide, circular basal depression which is about one third the diameter of the cup and is formed by the dorsal inflexion of the radials. Largest specimen attaining a width of 26 mm. at a height of about 13 mm. at the base of the tertibrachials. The theca possesses uninterrupted pentamerous symmetry. The surface is smooth and the sutures are not incised.

Basals.—These consist of four small equal plates grouped in a rounded cap about 3·5 mm. in diameter over the basal depression. For their lower few mm. they form a slightly expanding cyclindrical cavity which envelops a cylindrical calcite plug which in turn invests the proximal portion of the stem. They rest on the upturned, slightly bevelled ends of the radials. The five-petalled axial canal of the stem is continued through the basals into the thecal cavity. Inside the theca



TEXT-FIG. 5.—*Eucalyptocrinites inchoatus* sp. nov. Showing the arrangement of brachials within the arms, and the disarticulated brachials and radial partitions in the matrix above the cup. Note that one iBr is continued without interruption to the top of the specimen; both from M.U.G.D. 3405. × 2.

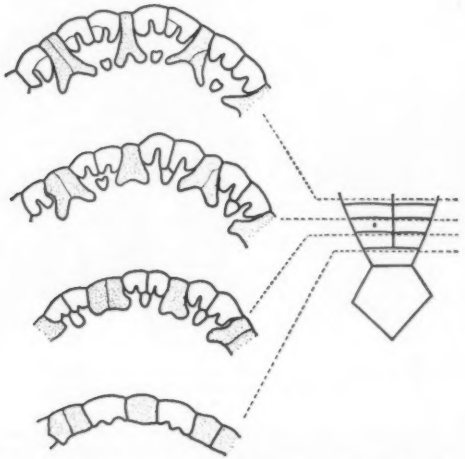
the basals reach at least as high as the $IIBrr_2$. Toward its apex the circlet of basals bears four external sutural ridges (Text-fig. 7j).

Radials.—These are five strongly inflected plates which circle the basal depression of the cup and are in contact with their adjacent radials for about 1·5 mm. on the outside of the cup. The externally visible portion of each radial typically extends 2 mm. up the cup and has a width of 6 mm. Above they support the quadrangular IBr_1 on a concave facet, while dorso-laterally, together with an adjacent radial, they support the large, lower iIBrr.

Primibrachials.—IBr_1 is a small quadrangular plate (h = 3 mm.; w = 4·5 mm.) with convex lower and lateral sutures, bearing a larger hexagonal IBr_2 (= IAx) (h = 3·5 mm.; w = 5 mm.), which, on its upper suture, supports the high iIIBr, while dorso-laterally it bears the asymmetric $IIBrr_1$.

Secundibrachials.—$IIBr_1$ is an irregularly pentagonal plate (h = 3·5 mm.; w = 3·5 mm.) supporting the low $IIBr_2$ (= IIAx) (h = 1·5 mm.; w = 3·5 mm.) which is the last plate of the ray system flush with the surface of the cup, as the rays above this level become progressively more raised above the iBrr.

Tertibrachials.—The $IIBr_2$ bears a series of IIIBrr which become progressively lower and wider up the arms ($IIIBr_1$, h = 1 mm.; w = 2·5 mm.: $IIIBr_4$, h = ·7 mm.; w = 3 mm.). In each arm there are three to five IIIBrr before the arms become biserial. The arms were probably free above the $IIIBrr_1$. In cross-section the lower IIIBrr are " U "-shaped with an open brachial groove (Text-fig. 6).



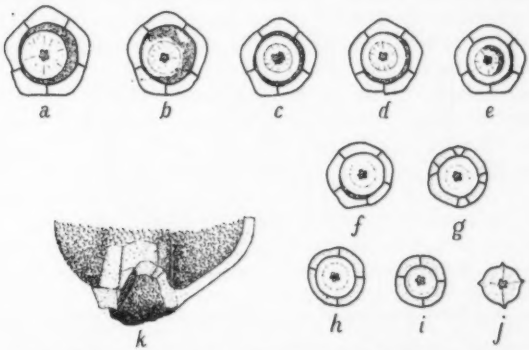
TEXT-FIG. 6.—Transverse sections at various levels through the lower part of the arms of *Eucalyptocrinites inchoatus* sp. nov. Showing the " Y "-shaped cross-section of the interbrachials above the IIBrr, and the cylindrical rod arising from the junction of the $IIIBrr_1$. Brachials white, interbrachials stippled. From S.M. H7524; sections × 2.

Interbrachials.—The circular ten-sided lower iIBrr are the largest cup plates (h = 6 mm.; w = 5·5 mm.). Each is supported by two RR while each in turn bears an upper elongate interradial area on its two upper sutures. These latter consist of pairs of plates (two iIIBrr [1]) separated by a vertical suture which swings around to one or other side of the area to occlude one of the plates, usually below the level of

[1] The existing plate nomenclature is ambiguous at this point for an iIIBr interradial in position is not distinguished from an iIIBr within a ray.

the IIIBrr$_4$. The remaining plate then extends well up to the crown forming a partition separating the arms at the surface of the cup.

Within each ray the only interbrachial present is an iIIBr which is similar in shape to the interradial iIIBr areas, but slightly narrower (w = 3 mm. as opposed to 4·5 mm.). Like the interradial iIIBrr, they are " Y "-shaped in transverse section above the IIIBrr, with one branch extending behind each arm pair (Text-fig. 6). No specimen



TEXT-FIG. 7.—a–j, Transverse serial sections of the basal depression of *Eucalyptocrinites inchoatus* sp. nov., from M.U.G.D. 3405. Distances of each section inside the basal depression: a, 1·3 mm.; b, 2·0 mm.; c, 2·7 mm.; d, 3·3 mm.; e, 3·8 mm.; f, 4·3 mm.; g, 5·0 mm.; h, 5·3 mm.; i, 6·3 mm.; j, 8·0 mm. Radials and cylindrical plug white; stem with radial lines where it can be clearly differentiated from the cylindrical plug; basals lightly stippled; matrix heavily stippled. k, Specimen broken to show the strongly inflected radials giving the basal depression. M.U.G.D. 3406. The basals are missing in this specimen. All × 1½.

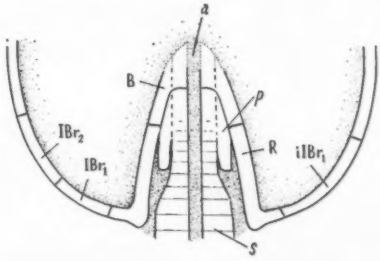
shows the junction of these plates behind the arms, or their imbrication with other plates to form the tegmen, but they appear to maintain their " Y "-shape for most, if not all, the distance to the top of the crown. Arising at the junction of the IIIBrr$_1$ in each arm pair is a small, cylindrical vertical rod which becomes free of the brachial system toward the top of the IIIBrr$_2$, and then is located between the ends of the " Y "-shaped branches of the iIIBrr at least as high as the IIIBrr$_4$.

The relation of these plates to the tegmen proper is not known, but they probably imbricated with the lower tegmenal plates.

Columnals.—Columnals associated with the cups are cylindrical, disc-shaped plates reaching 6 mm. in diameter and a thickness of 1 mm. They possess a strongly petalloid axial canal with marked pentagonal symmetry. Their articulating surfaces are marked by strong radiating

ridges. None of the stem fragments is differentiated into nodals and internodals. The stem is apparently anchylosed to the radials and basals inside the basal depression of the cup by a cylindrical calcite plug enveloping the proximal portion of the stem. Longitudinal sections of stem fragments which probably belong to this species are discussed on p. 156 (also Text-fig. 9).

Remarks.—Because the deep basal depression of the cup was found impossible to excavate, two specimens were serially sectioned to show



TEXT-FIG. 8.—Reconstruction of a longitudinal section of the basal part of the cup of *Eucalyptocrinites inchoatus* sp. nov. Based on the transverse sections shown in Text-figs. 7a–j. The sutures between the columnals are schematic, and the shape of the upper part of the basals is inferred. Matrix stippled; the vestigial sutures between the proximal portion of the stem and the calcite plug and the trace of the stem outline within the basals shown as a broken line; *a*: axial canal of stem; *p*: calcite plug; *s*: stem. × 2.

the basal plates. The basals had been lost in one of these specimens. However, beside revealing the basals, this procedure also provided some information as to the stem attachment in *E. inchoatus* (Text-fig. 7, 8).

At and immediately inside the basal depression the stem is much smaller than the inside diameter of the depression which is determined by the circlet of upwardly inflected radials (Text-fig. 7a). Between a distance of 1·3 mm. and 2 mm. inside the depression, the stem contracts rapidly to less than three-quarters its diameter and is here first surrounded by a cylindrical calcite plug which envelops the stem above this level. The stem and plug remain roughly at the same diameter for the next 3 mm. while the circlet of radials, forming a cone, gradually contracts around the stem and plug, till, at about 2·5 mm. after the plug first appears, it becomes firmly anchylosed within the radials. Below this level the plug is separated from the radials by matrix. At about this level the basals first appear as four equal plates resting on the slightly bevelled surface of the radials. Above their appearance the four

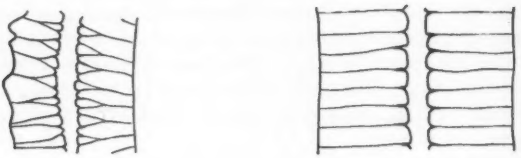
basals take the place of the radial circlet, surrounding the calcite plug for the next few millimeters. The stem and plug are no longer clearly separable above the last stem joint encountered at 1·8 mm. above the base of the plug (Text-fig. 7e). From here upwards the junction is present only as a slightly visible discontinuity right through into the cap of basals. After the basals appear they continue the contraction of the conical depression at the expense of the outer part of the plug, while the vestigial outline of the stem within the plug remains unaltered in diameter. The basals swing over to cap the depression when the trace of the outer part of the calcite plug has been excluded by the tapering circlet formed by their lower part. The trace of the stem outline, seen in section as a darker-coloured zone of the calcite, passes up into the upper part of the basals, which, at 8 mm. from the base, have developed outer, sutural ridges. The petalloid axial canal of the columnals passes up through the basals apparently into the theca without interruption or change in diameter.

The calcite plug is considered to be a primary structure rather than an effect of fossilization, first because of its relation to the matrix which suggests it must have been in position before burial, secondly because of its very regular shape, and finally because of its apparently intimate association with the upper part of the stem. It is unfortunate that the structure is known only from one specimen and similar structures do not appear to have been reported previously.

Apparently the plug would have served to anchor the stem to the cup, but its effect on the growth of the stem is difficult to interpret. If the plug were anchylosed to the distal portion of the stem as the sections suggest then presumably it would have prevented introduction of new columnals beneath the basals. However there is no evidence for considering that the stem grew in a manner other than this; there is no well-developed columnal at the proximal end of the stem which could have served a function similar to that of the proximale of articulate crinoids. A longitudinal section of the basal part of a cup in which the stem and plug were *in situ* would throw more light on the problem, particularly as it would establish the presence or absence of sutures between the upper columnals. In the transverse sections no trace of these horizontal sutures were seen in the stem above 1·8 mm. from the base (Text-fig. 7e), but they are extremely difficult to detect in sections parallel to them.

In fact there appears to be very little information on the mode of attachment of stem and cup and the growth of the stems in Palaeozoic crinoids. Wachsmuth and Springer (1897) maintained that in all camerates and inadunates new columnals were introduced at the base of the cup so that the top columnal is always the last formed. Of this view, however, Bather (1898, p. 324) has commented that " so slight is

the attempt at proof that scepticism remains our only attitude". Wachsmuth and Springer's generalization does not hold good for all camerates. In such monocyclic genera as *Polypeltes* the stem is attached to the cup in a shallow depression which may contain up to fifteen of the lower cup plates. Moreover, during ontogeny, the shallow depression encroaches further up the cup to include more plates and also the basals may be partially resorbed (*fide* Ubaghs, 1956b). Hence in this case the upper columnal is a plate apparently homologous to the proximale of articulate crinoids.



TEXT-FIG. 9.—Longitudinal sections of stem fragments of *Eucalyptocrinites inchoatus* sp. nov. a, from M.U.G.D. 3407; b, from M.U.G.D. 3409 B. × 4.

The columnals of *E. inchoatus* are externally similar to those illustrated for other species of *Eucalyptocrinites*, although certain species (e.g. *E. ovalis* (Hall), Wachsmuth and Springer, 1897, pl. 82, fig. 2; *E. pernodusus* Springer 1926, pl. 9, fig. 5) have a stem differentiated into nodals and internodals. A number of longitudinal sections was ground of various stem fragments which were identified as belonging to this species although they were not directly associated with cups. In the smallest and presumably most immature of these (Text-fig. 9a) the columnals can be seen to consist of two alternate series—a series of larger ossicles extending from the axial canal to the stem surface and thickening away from the axial canal, and a smaller series which tends to decrease in size away from the axial canal so that in general the ossicles of this series are occluded by those of the larger series and do not reach the stem surface. In larger fragments (Text-fig. 9b) which were presumably more mature, this dimorphism is completely lost and the stem is made up of flat similar cylindrical ossicles. A similar, although more complex, arrangement of ossicles has been described by Ubaghs (1956a) in *Barrandeocrinus sceptrum* and is probably widespread throughout the camerate crinoids. The irregular outer surface of the stem illustrated in Text-fig. 9a suggests that the specimen was pathological or damaged in some way before burial in the limestone.

A further feature of morphological interest in this species is the presence of the cylindrical rods arising from inside the junction of the $IIIBr_1$ within each arm pair (Text-fig. 6). Clearly these would have served to separate the food currents from each arm and form separate passages into the tegmen. It is doubtful whether the arms could have been free till these rods were separated from the brachials, i.e. till above the $IIIBrr_2$. Similar structures have been illustrated in other species of *Eucalyptocrinites* (e.g. *Hypanthocrinus granulatus* Lewis, Angelin, 1878, pl. 24, fig. 10 and perhaps *Eucalyptocrinus ovalis* Hall, Wachsmuth and Springer, 1897, pl. 82, fig. 3), but it is difficult to decide whether the type species *E. rosaceus* Goldfuss has this character or possesses but one tegmenal opening for each arm pair. The usually reproduced figure of Schultze (1866, pl. 11, fig. 10) shows this latter situation, but his fig. 8 shows single openings for each arm.

One character which distinguishes *E. inchoatus* from the northern hemisphere members of the genus is the apparent absence of extra plates in the interbrachial partitions. In *E. inchoatus*, as far as can be judged, the partitions consist solely of an iIIBr. In other species the radial partitions are separate plates horizontally supported by the interbrachials usually at a height between the $IIIBrr_3$ and the $IIIBrr_5$. In most Silurian species these partitions then extend well up to the top of the crown as in *E. regularis* (Hisinger) (Angelin, 1878, pl. 24, fig. 11) or *E. decorus* (Phillips), the type species of *Hypanthocrinus* Phillips, 1839, or they may be horizontally surmounted by a partitional plate of similar size as in *E. rosaceus* Goldfuss (Schultze, 1866, pl. VIII, fig. 4).

Superfamily HEXACRINITICAE Ubaghs
Family HEXACRINITIDAE Wachsmuth and Springer
Genus **Hexacrinites** Austin and Austin, 1843
Hexacrinites (?) sp.
(Text-fig. 10; Pl. VIII, figs. 5–7)

Material.—One fragmentary internal mould of a cup with portion of its counterpart (M.U.G.D. 3411).

Description.—The cup is relatively high and cone-shaped with a markedly hexagonal cross-section. The internal mould reaches a height of 14 mm. and has a similar width opposite the top of the radials.

Basals.—These consist of three large equal triangular plates (h = 8 mm.; w = 9·5 mm.) extending well up the cup. Medially they support a radial on a horizontal, sagging suture, while on each side of their dorsal margin they share another radial with the neighbouring basal, or posteriorly the interradial.

Radials.—These consist of five subequal plates, the anterior three of which average 11·5 mm. in height and 8 mm. in width. They rest on the basals and form a circlet of plates interrupted posteriorly by a

similarly shaped interradial. The posterior radials are slightly narrower than the anterior radials (w = 6·5 mm.). All the radials are basically quadrangular in shape, with a deep triangular notch at the centre of their upper edges, apparently to support the brachials. The radials are constricted by a shallow furrow immediately beneath this notch.

Posterior Interradius.—This is marked by a single plate incorporated in the circle of radials. It is similar in shape to the radials, and comparable in width to the anterior ones, so it is slightly wider than the neighbouring two posterior radials. It is supported by the dorso-lateral facets of the two posterio-lateral basals.

TEXT-FIG. 10.—*Hexacrinites*(?) sp. Inferred plating; basals white; radials black; posterior interradial stippled. × 1½.

Brachials, pinnules, tegmen, stem, and the surface of the cup are not known.

Remarks.—The single fragmentary specimen can only be questionably referred to the genus *Hexacrinites*, particularly as the surface features of the cup are not known. Although the forty-nine species of *Hexacrinites* listed by Bassler and Moodey (1943) are exclusively Devonian, Strimple (1952) has described a Silurian species from the Henryhouse Formation of Oklahoma.

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I am indebted to Mr. A. G. Brighton, of the Sedgwick Museum, Cambridge, who very kindly read and criticized the manuscript of this paper, and to Prof. G. Ubaghs of Liège University who commented on the specimens during a brief visit to Cambridge. The manuscript was prepared during the tenure of an Australian Commonwealth Scientific and Research Organization Overseas Studentship.

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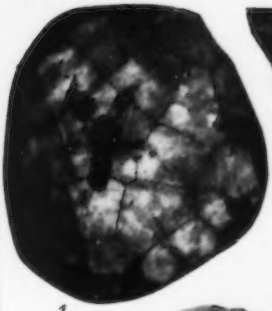
EXPLANATION OF PLATE VIII

ALL FIGS.— × 2 unless otherwise stated; Figs. 1–7 retouched.

FIGS. 1, 2, 8.—*Thylacocrinus*(?) *ignotus* sp. nov. 1, Holotype, M.U.G.D. 3400, internal view of base of cup; 2, Lateral view of internal mould of cup unconventionally lit to show the plate impressions, M.U.G.D. 3401. × 1·5; 8, Internal mould of base of cup, M.U.G.D. 3402.

FIGS. 3, 4, 9.—*Eucalyptocrinites inchoatus* sp. nov. 3, Basal view, S.M. H. 7524; 4, Lateral view of holotype, M.U.G.D. 3404; 9, Articulating surface of a columnal, M.U.G.D. 3407. × 5.

FIGS. 5–7.—*Hexacrinites*(?) sp. 5, Basal view of internal mould, M.U.G.D. 3411A; 6, Lateral view of same specimen; 7, Lateral view of counterpart, M.U.G.D. 3411B.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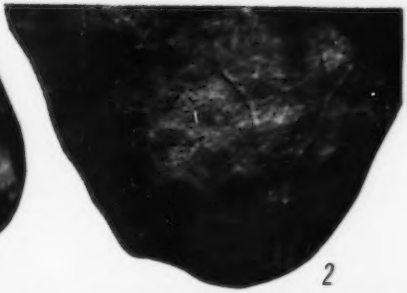
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LOWER DEVONIAN CRINOIDS FROM AUSTRALIA

The Upper Devonian and Lower Carboniferous Stratigraphy of Boscastle and Tintagel, Cornwall

By E. B. Selwood

Abstract

The discovery of trilobites characteristic of the *Gattendorfia* Zone is described from the California Quarry, Boscastle. The possible stratigraphical implications are discussed and a comparison is made with beds of similar age at Launceston. It is suggested that there is no important stratigraphical break at the top of the Devonian and that there is a conformable passage from the Tintagel Volcanic Series into the Carboniferous. The Tredorn Phyllites are tentatively equated with the Woolgarden Phyllites. There is no evidence to support the contention that the Lower Culm contains an important greywacke sequence; these beds are probably of Upper Culm age.

Introduction

Although successions have been established for many years in beds considered to be of Upper Devonian and Lower Carboniferous age in North Cornwall, the nature of the Carboniferous–Devonian boundary remained obscure. Recently however, the author has shown (1960), on the basis of the faunas, a conformable sequence of strata between the two systems at Launceston. A search for these faunas and the associated lithologies on the coast some 15 miles to the west, in the region of Tintagel has yielded the results set out in this paper.

A review of the history of research on the Upper Devonian strata of the region together with concise descriptions of the lithologies has been given by Wilson (1951). The classification in general use today is that of Dewey (1909), but recently Batstone (1959) has proposed that the Barras Nose and Trambley Cove Beds should be included in a more widely defined Tintagel Volcanic Series (Table 1).

Table 1

Dewey 1909	Batstone 1959
Tredorn Phyllites	Upper Grey Slate Series
	——— Structural Break ———
Trambley Cove Beds	Tintagel Volcanic Series: Upper Black Slates; Trewethet Volcanics
Volcanic Series	Tintagel Volcanic Series: Lower Black Slates; Davidstow Volcanics
Barras Nose Beds	Tintagel Volcanic Series: Lower Black Slates; Davidstow Volcanics
Woolgarden Phyllites	Woolgarden Series
Delabole Slates	Delabole Slate Series
	——— Structural Break ———
Slaughterbridge Beds	Slaughterbridge Series

Workers are agreed that the junction between the Devonian and Carboniferous systems appears on the cliffs below the California Quarry, Boscastle, but the exact nature of the contact has been

disputed. Wilson (1951) has observed that "there is no apparent structural break between the dark grey slates and grits of the Culm above and the pale grey-green phyllites below, but the contact is generally recognized as an unconformity". Owen (1950) on the other hand believes that the contact is conformable, whilst Dewy (1948) interprets it as a fault.

The details of the Carboniferous succession on the North Cornwall coast have been described by Owen (1950) who recognizes within the Lower Carboniferous the following conformable sequence:—

4. Cleaved slates on Rusey Cliff
3. Slates with thin grit bands on Buccator Cliff
2. Cherts at Firebeacon Point
1. Shales with thin grit bands at Boscastle.

In an earlier paper Owen (1934) referred this part of the Carboniferous succession to the Lower Culm rather than to the Lower Carboniferous. This earlier usage is preferred by the Geological Survey (Dewey 1948).

In a work embracing the whole of the Culm succession on the North Cornish coast, Ashwin (1958) refers Owen's Lower Culm to the Boscastle Measures. In the absence of fossils this terminology was to be preferred.

The California Quarry

The beds which structurally overlie the Tredorn Phyllites at the California Quarry are medium dark-grey, spotted phyllites which bear abundant evidence of movement along planes more or less parallel to the bedding. A similar lithology is exposed in a small quarry a few yards to the north (Grid ref. SX 092 909), but here the phyllites contain occasional thin siliceous seams and appear to pass conformably upwards into dark phyllites with numerous thin siliceous seams. These seams are very fine-grained and probably represent thin cherts rather than fine sandstones such as those which characterize the Culm rocks around Boscastle Harbour.

Hitherto the dark phyllites of the California Quarry have yielded no fossils, but the author has obtained a sparce fauna which includes crinoid ossicles, brachiopods, lamellibranchs, polyzoa, solitary corals and trilobites. The fossils are highly distorted, but it has been possible to refer some of the trilobites to the subgenus *Cyrtosymbole* (*Macrobole*) and others to *Cyrtosymbole* (*Waribole*). Whilst species of *C.* (*Waribole*) range from the Upper Devonian into the Lower Carboniferous, species referred without reservation to *C.* (*Macrobole*) are exclusively Tournaisian in age. From cranidia (GSM 87364-70) and pygidia (GSM 87371-2) it has been possible to recognize the species *C.*

(*Macrobole*) cf. *drewerensis* Rud. and E. Richter and *C.* (*Macrobole*) cf. *duodecimae* Rud. and E. Richter. These species, originally described from the Sauerland, Germany, have been discovered by the author in the Launceston area and *C.* (*Macrobole*) *duodecimae* has been described by Goldring (1955) from North Devon. In all cases the species have been associated with the goniatite *Gattendorfia ;* the index fossil for the lowest zone of the Carboniferous. It is generally considered that these species are restricted to the *Gattendorfia* Zone. Thus although goniatites have not yet been found in them, the dark phyllites may be referred to the *Gattendorfia* Zone with some confidence.

Since the Tredorn Phyllites are undoubtedly Devonian the junction between them and the dark phyllites is indeed a junction between the Devonian and Carboniferous systems. However, the author agrees with Dewey (1911) that the Tredorn Phyllites are not the highest strata of the Devonian; it is therefore probable that this junction is a fault. It will be argued below that the slates with thin siliceous seams which structurally overlie the dark phyllites belong to the *Wocklumeria* Zone. Thus the system boundary is here present in an uninterupted, but inverted succession, and the fault beneath the dark phyllites is confirmed (Table 2).

TABLE 2.—THE SUCCESSION AT THE CALIFORNIA QUARRY

4	Greywackes	Upper Culm
	——Fault——	
3	Dark siliceous phyllites with an interbedded lava	Upper Devonian (?*Wocklumeria* Zone)
2	Dark fossiliferous phyllites	Lower Carboniferous (*Gattendorfia* Zone)
	——Fault——	
1	Tredorn Phyllites	Upper Devonian (?*Clymenia* Zone)

THE DEVONIAN

The succession within the Upper Devonian according to Parkinson and Dewey was based on the assumption of a conformable sequence of strata with a regional dip to the north. However, Batstone (1959) has maintained that there is a structural break at the base of the Tredorn Phyllites. The field evidence is not easy to interpret but the author would agree with Batstone. Generally where seen the contact is obviously faulted.

Though undoubtedly Upper Devonian the position of the Tredorn Phyllites within the succession now becomes doubtful; particularly as it is generally agreed that the Woolgarden Phyllites and Delabole Slates successively underlie the Barras Nose Beds. It is often difficult to distinguish the Woolgarden Phyllites from the Tredorn Phyllites,

for the banding which is described as characteristic of the Woolgarden Phyllites is not always developed. Both phyllites have a similar fauna yielding locally abundant but highly distorted spirifers, rhynchonellids, crinoids and polyzoa. The similarity of the lithology and fauna suggests that both are representatives of the same unit. If this is so then the Barras Nose Beds are the highest Upper Devonian rocks present in the Tintagel area.

The dark phyllites with thin siliceous seams of the California Quarry are lithologically indistinguishable from the Barras Nose Beds and the Trambley Cove Beds. This, together with the presence of an inter-bedded vesicular lava in the chasm of Western Blackpit (Willapark) is strong presumptive evidence for equating them with the Tintagel Volcanic Series as defined by Batstone.

The Carboniferous

The beds between Boscastle and the Rusey Fault, which have usually been referred to the Lower Culm, show a complex tectonic pattern which involves cross folding (Ashwin, 1957). Owen's succession, which was built up on the assumption of an orderly sequence of strata dipping to the north, therefore becomes suspect.

The oldest rocks of Carboniferous age are the trilobite-bearing phyllites at the California Quarry, but their relation to the rest of the Culm succession is obscured by the intervention of the siliceous slates and lava of the Tintagel Volcanic Series. The contact between the latter and the greywacke sequence which forms the bulk of Owen's Lower Culm succession is not exposed except in an inaccessible cliff, but the form of the cliff suggests the presence of a fault.

According to Owen the radiolarian cherts at Firebeacon Point are inter-bedded in the greywacke succession. However, close examination of the contact shows that the cherts are structurally discordant on the greywackes and in fact only cap the headland. The stratigraphical position of the greywackes is thus questionable, for they have yielded no fossils and are not observed in conformable contact with any dated horizon.

It has been alleged that limestones are absent from the Lower Culm (Owen, 1950). But faulted within the greywackes between Firebeacon Point and the Rusey Fault are slates bearing thin limestones. At Gull Rock (Grid ref. SX 117 933) these have yielded goniatites which have been tentatively referred by Dr. J. E. Prentice (personal communication) to the P1 Zone of the Lower Carboniferous.

There are therefore only two horizons which can with certainty be referred to the Lower Culm; namely the fossiliferous dark phyllites and the limestone horizon, neither of these has been recorded before. Thus two major stratigraphic units, the cherts and the greywackes are

unplaced stratigraphically. Their correct position can only be assessed at present by reference to adjoining areas (Table 3).

Comparison with other Areas

Table 3.

	Boscastle-Tintagel	Launceston
Upper Culm	Greywacke Beds	Trebursye Beds
Lower Culm	Slates with Limestone Firebeacon Cherts Dark Phyllites	Cannapark Limestone and Barracados Cherts Yeolmbridge Beds
Upper Devonian	Volcanic Series Woolgarden Phyllites (Tredorn Phyllites) Delabole Slate	Stourscombe Beds Petherwin Beds

The correlation between the coastal and Launceston areas suggested above is based largely upon lithological similarities, but the occurrence of index fossils at two horizons on the coast does provide a certain amount of control. The key to the correlation proposed lies in the existence of a *Gattendorfia* Zone fauna at the California Quarry which allows the dark phyllites from this quarry to be correlated directly with the Yeolmbridge Beds at Launceston.

At Launceston, beds of *Wocklumeria* age, the Stourscombe Beds, have been proved faunally; these beds are composed of black slates bearing thin chert bands and nodules and include an important volcanic horizon. It was formerly supposed (Reid *et al* 1911) that this lava was interbedded with Culm slates and sandstones. On this basis Owen (1950) correlated it with the lava at Willapark which he believed, mistakenly, to be interbedded with a similar lithology. The correlation is nevertheless probably still valid, for the Willapark lava is in fact interbedded with black slates containing thin cherty bands; a lithology which in both areas lies in comfortable contact with rocks of *Gattendorfia* age.

Thus the Volcanic Series of Tintagel is correlated with the Stourscombe Beds of *Wocklumeria* age and is followed by beds with a *Gattendorfia* Zone fauna which correlate with the Yeolmbridge Beds. The Volcanic Series is underlain by Woolgarden Phyllites, and these, allowing for the higher metamorphism, are comparable with the Petherwin Beds (*Clymenia* and *Platyclymenia* Zones) which underlie the Stourscombe Beds at Launceston. The calcareous development of the Petherwin Beds cannot be recognized on the coast, but this is not unexpected for the limestone is very localized even around Launceston. In so far as it is possible to judge, (the deformation making specific identification impossible) the general make-up of the

fauna from the Woolgarden Phyllites and Tredorn Phyllites is closely comparable with that from the upper part of the Petherwin Beds.

Although the Firebeacon cherts have yielded no macrofossils they resemble the Barracados cherts lithologically and have long been considered to be their equivalents for it is possible to trace the horizon, more or less continuously, from Firebeacon Point to an area east of Launceston (Owen 1950). Ussher believed that the main limestone development in the Lower Culm was younger than the radiolarian chert; it has not been possible to prove this in North Cornwall, but there is no indication that the limestone is older than the *Goniatites* Zone.

The inclusion of an important series of greywackes in the previously accepted Lower Culm succession of North Cornwall has always created an anomalous situation in the stratigraphy of the South-West, for comparable beds elsewhere are younger and are referred to the Upper Culm. Ussher (1901) in fact emphasized that the Lower Culm contains no grits. It has been pointed out (p. 164) that the greywacke-bearing strata on the coast cannot be directly placed stratigraphically, since they are neither in conformable contact with beds of Devonian age nor with the radiolarian cherts of Firebeacon Point. Thus whilst it has not been possible to prove Ussher's contention by fossils either north of Boscastle or at Launceston, it has been possible to demonstrate first at Launceston (Selwood, 1958) and now on the coast that the greywackes are not interbedded with the strata considered to be of Lower Culm age. In the absence of any evidence to the contrary these greywackes are here removed from the Lower Culm and referred to the Upper Culm.

Acknowledgements

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DEPARTMENT OF GEOLOGY,
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CORRESPONDENCE

ORIGIN OF CONVOLUTED LAMINAE

SIR,—A serious flaw in J. E. Sanders's interpretation of the origin of convolute bedding (*Geol. Mag.*, 1960, xcvii, 409–421) is implicit in the last few lines of his paper. He writes that " the phenomenon is known only from alleged turbidite sequences in the geologic record ", yet he recognizes that the velocity fluctuations which his interpretation requires are not believed to be associated with turbidity currents.

We can regard the normal unit of the turbidite sequence as having been deposited relatively quickly *as a whole*. In such a deposit it is most difficult to imagine convolute " folds " (which may be within, and some distance from, both top and bottom of such a unit) as being produced in the form of surface relief as " streaked-out ripples " by a fluctuating water current passing over a silty bottom deposit.

Of greater importance is that any theory of the origin of convolute bedding should *positively take account of* the association of the phenomenon with turbidite sequences. The mechanism suggested by Sanders could, and presumably would, operate equally in any other situation, provided that the bottom deposits were of the critical grade and that the appropriate fluctuating currents were present.

The particular feature of turbidite deposition that should be accounted for in any theory of the origin of convolute bedding has already been mentioned. It is that such deposits are laid down relatively quickly and as substantial units. Thus a comparatively large thickness of unconsolidated and water covered sediment is available *at one time*: perhaps for the operation of some process involving localized liquifaction and hence flow of the sediment. A promising explanation on these lines has recently been given by Emyr Williams in his paper on " Intra-Stratal Flow and Convolute Folding " (*Geol. Mag.*, 1960, xcvii, 208–14), and more especially in an early draft of his subsequent letter on this subject (below) where he refers to the work of Terzaghi.

DEPARTMENT OF GEOLOGY, C. H. HOLLAND.
BEDFORD COLLEGE,
LONDON, N.W. 1.
21st November, 1960.

SIR,—It was with much interest that I read the stimulating new ideas put forward by Dr. J. E. Sanders to account for convoluted laminae (*Geol. Mag.*, xcvii, 409–421). However, the mechanism proposed for the contortions is unacceptable for a number of reasons. It is stated that this mechanism is met only if the affected water-saturated and originally cohesionless sand can be alternately cohesive and cohesionless (p. 419), but according to the experimental evidence known to the writer, such behaviour is unknown and does not appear probable. Again, it is pointed out that although convoluted laminae have been described only in alleged turbidites, the distortions according to this mechanism are anomalous when considering the theory of turbidity currents (p. 419)—which is based on much experimental work. Furthermore, convolute folding is typically characterized by very irregular attitudes of axial planes and directions of recumbency of the folds, which would be unlikely if the anticlines developed in a way similar to that of current ripples, as suggested by Sanders.

The occurrence in a number of examples of unconformities developed during the deformation causing convolutions is rightly stressed in Sanders' paper. In a recent paper (*Geol. Mag.*, xcvii, 208–14) I was in error in not realizing their importance, for they must indicate, when they are present, that the deformation occurred not within beds confined between others, but within water-covered beds.

Also of importance is that convolute folding is limited to sediment mainly of fine-grained sand and coarse silt, which is of the size-grade range susceptible to liquefaction. This fact coupled with the experimental observation that the structure of fine-sediment collapses quickly after deposition from turbidity currents (Dzulynski, Ksiazkiewiecz, and Kuenen, 1959, *Bull. Geol. Soc. Amer.*, lxx, 1112), thus showing its initial metastable nature, indicates that it is necessary to reckon with the phenomenon of liquefaction.

Liquefaction is best understood by examining the shearing resistance equation of water-saturated unconsolidated sediment, and since, according to normal usage in soil mechanics, I misused the terms "effective" and "neutral" in the explanation given in my paper (p. 208) it is appropriate to reconsider this equation:—

$$s = c - (p - u) \tan \phi$$

where s = shearing strength; c = cohesion; $p = \gamma_s z_s$, wherein γ_s is submerged unit weight of sediment and z_s depth below free surface of sediment; u is pore water pressure in excess of value of $\gamma_w z_w$, wherein γ_w is unit weight of water and z_w depth below water surface; ϕ = angle of internal friction. Normally "u" is zero. However, if a metastable structure in a granular, cohesionless ($c = o$) sediment collapses, the term "u" increases almost immediately from zero to a value which may be close to the pressure "p", with the result that the shearing strength of the metastable deposit drops to a value which may be close to zero, when the sediment temporarily behaves as a liquid (see Terzaghi, 1957, Teknisk Ukeblad Oslo, nr. 13–14, 1–3).

Terzaghi (*ibid.*) discusses examples of the liquefaction of fine-grained and cohesionless sediment in terms of impulses received from earthquakes, blasting operations, and seepage pressures exerted by the flow of ground water towards the ocean during very low tides, etc. Of particular interest are his conclusions regarding the liquefaction of water-covered sediment, which may be resting on flat surfaces. It is maintained that liquefaction commonly starts at a point and spreads over an increasingly larger area. However, because sedimentary deposits with an unstable texture usually contain layers and lenses of material with stable texture, liquefaction proceeds erratically, and there are relative displacements within the agitated sand. The mechanics of progressive liquefaction are explained by considering that some of the excess water expelled from the temporarily liquefied part percolates laterally causing liquefaction of the intact portion.

According to Terzaghi's mechanism for progressive liquefaction it seems that contortion of laminae must result from the displacements within affected water-covered, fine-grained, unstable, and cohesionless sands, even when they are reposing on flat surfaces. Due to the lateral percolation of some of the expelled water from liquefied parts, the development of the folds would closely follow the patterns given in my paper, and the deformation would die out against an approximately smooth bottom. It is suggested that the topmost material, as a result of agitation with the escape of part of the expelled water upwards, resettles giving a smooth surface, so that the folds within the affected layer would, in general, appear to die out against the upper margin, although slight unconformities would be expected.

In a private discussion during July–August, 1960, Dr. C. H. Holland implied that none of the causes listed in my paper for intra-stratal flow within confined layers satisfactorily explained the convolute folding occurring with random orientation within medium-coarse calcareous quartz siltstones, which are considered to be turbidites, in the Lower Ludlovian rocks of North-East Radnorshire (Holland, 1959, *Geol. Mag.*, xcvi, 230–6). He wondered if the mechanism proposed could be applied to the sedimentary layer immediately after deposition. On my suggesting that in these rocks convolute folding may have resulted from the liquefaction of unconsolidated water-covered beds as described above, since some of the crests of the folds are cut off by succeeding laminae (p. 231), Holland stated his support.

It appears that liquefaction would occur in water-covered sediment with a much narrower size-grade range of material and a more unstable texture

than in water-logged confined layers, from which water would not be immediately expelled during disturbance. It is also considered likely that the directions of the convolute fold axes, which according to my mechanism depend to a large degree on the distribution of the pockets of material that had retained a notable shearing strength, would be more haphazard resulting from the liquefaction of water-covered deposits than in examples due to directed movements within confined layers.

Finally, it may be noted from the foregoing that the reasons which make Sanders' mechanism unacceptable are satisfied by the mechanism proposed by the writer. In addition, exaggeration, distortion, and overturning of primary upward protuberances, and the inversion of the filling of depositional depressions, as well as the irregular attitudes of the axial planes and directions of recumbency of the folds, can be adequately explained by laminar flow within liquefied layers, whether confined or, in the examples where unconformities developed during deformation, water-covered.

GEOLOGICAL SURVEY DEPARTMENT, E. WILLIAMS.
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1st January, 1961.

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THE PATTERN OF SOME PACIFIC ISLAND CHAINS

SIR,—The article entitled, " The Pattern of Some Pacific Island Chains," by L. J. Chubb (1957) has just reached the present author. Chubb's main hypothesis deals with the evolution of the island chains and origin of the coral reefs as follows: " Nearly all the islands discussed in this paper must have originated in pre-glacial times, so they would be expected to furnish evidence of any post-glacial rise in sea-level. But this rise would have been everywhere equal, the sea cannot have risen by different amounts around different islands, yet the degree of drowning of the coasts and the stage in the development of reefs varies from island to island within any one chain. The only explanation of this fact is that *each island subsided independently*, . . ." (italics supplied).

The writer, during nearly two decades as District Geologist in the Hawaiian Islands and as Geologist-in-Charge of Pacific Ocean investigations for the Geological Survey, has never found any fact to substantiate Mr. Chubb's conclusion. For example, on page 222 he states, " The Hawaiian chain shows a very complete sequence but it is characterized by the *absence of drowned river-valleys* and of barrier reefs, though fringing-reefs and atolls occur " (italics supplied). All of the islands where not veneered with recent lava have deeply drowned river valleys which have been described in the literature since 1935 (Stearns 1935a, 1935b, 1935c, 1946a, and Stearns and Vaksvik 1935). Logs of wells at the mouths of these valleys indicate that the subsidence amounts to more than 1,200 feet.

In the same paragraph on page 222 Chubb states: " Eastward lies Kauai, an ancient volcano so deeply dissected that all trace of its original form is lost." Kauai contains near its centre a lava filled caldera 10 miles across and although it is deeply eroded there is no difficulty in tracing its original form (Stearns 1946a, Macdonald, Davis, and Cox 1960).

On page 223 Chubb states: " The only important embayment within the chain, Pearl Harbour in Oahu, is due, not to the drowning of a river-valley, but to the depression between the two volcanoes composing the island being still below the sea. This need have involved neither subsidence of the island nor a rise of sea-level." Pearl Harbour is due to the drowning of river-valleys (Stearns 1935a and b, and Stearns and Vaksvik 1935). The contact of the Koolau and Waianae lavas lies several miles west of Pearl Harbour.

On the same page he states that Tau, Olosenga, Ofu and Tutuila Islands in

the Samoan chain have no recognizable craters and the Pago Pago, the one important embayment on Tutuila resembles Pearl Harbour. The harbour of Pago Pago is a drowned river-valley along a large caldera escarpment. A caldera exists on Tau. Olosega and Ofu are erosion remnants of a single volcano which had a large caldera (Stearns 1944).

On page 224 is a discussion of the Cook group in which reefs 20 to 200 feet above sea-level are cited as evidence of uplift. Although some of these reefs may be due to uplift, many of the lower ones are undoubtedly due to eustatic shifts of sea-level in Pleistocene time in common with all other island groups of the Pacific that carry reefs and shore lines above sea-level (Stearns 1938a and b, 1941, 1945b).

We are led to believe from Chubb's paper that each volcano and island chain built up above sea-level by lava flows became extinct, was then eroded and finally subsided to develop first barrier reefs and then sink beneath the waves to become an atoll. This theory, " that each island subsided independently," is not supported by the facts so far as the present writer is aware in any of the island chains. The mechanism of independent subsidence of volcanoes built along either a single or double rift zone would require highly improbably geologic processes. Ample evidence exists that the Hawaiian archipelago, which is a good example of island chains, was built over two great rift zones with a migration of the volcanic activity lengthwise along both rifts ending with Hawaii at the south-east as the youngest island. Deeply drowned valleys on all of the islands, including Hawaii, and emerged shorelines as high as 1,200 feet above sea-level, indicate that the chain as a whole has survived as a unit mass over a long period of time, at least since the Pliocene.

The rate of weathering, which proceeds rapidly in the tropics, probably has been overestimated and it is possible that the Hawaiian Islands, where weathering extends downward more than 300 feet in some of the older rocks, may well indicate that the older volcanoes became extinct in early or middle Tertiary rather than in the late Tertiary as commonly believed. The presence of atolls at the north-east end of the chain with stacks and fringing reefs to the south-east can be explained better by the downward tilting and/or greater age of the chain to the north-west; and hence more erosion for the islands on the north-west end.

On page 228 Chubb states, " Dietz and Menard (1953) have described submerged terraces at various depths from 180 to 700 fathoms around several of the islands, which they interpret as drowned shorelines, indicating subsidence. But this is difficult to reconcile with the absence of drowned river-valleys, which suggests that the lack of barrier-reefs is due to the islands not having subsided." Here again Chubb was unable to reconcile the submerged terraces and the drowned shorelines. There is no doubt that submergence in the central Pacific has been of the order of 10,000 feet (Stearns 1946b).

All of the simatic islands in the central Pacific have subsided and at the same time have been subjected to eustatic shifts of sea-level (Stearns 1946c), the later ones due chiefly to glaciation and deglaciation in the order of 600 feet, ranging from about 300 feet below present sea-level to about 300 feet above present sea-level (Fairbridge 1960), while some of the islands beyond the andesite or sialic line have emerged, uplifting and tilting the reefs on which the Pleistocene eustatic shorelines were cut. In addition to the eustatic shorelines due to changes in the volume of the ice caps there is evidence of tremendous changes in the shape of the floor of the Pacific in Tertiary and Quaternary time which have left shorelines probably eustatic up to 1,200 feet above sea-level, not due to glaciation and deglaciation (Stearns 1945a).

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9th January, 1961.

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IRISH EROSION SURFACES

SIR,—Mr. E. H. Thornton has recently (1960) offered a statistician's comments on a published summary (Davies, 1958) of a paper on Irish erosion surfaces which I read to Section E of the British Association in 1957. Mr. Thornton reaches the conclusions, first that in the altimetric frequency graphs which formed the basis of my work the occurrence of peaks may be attributed largely to chance and that they do not therefore represent erosion surfaces, and secondly that the coincidence of peaks in the graphs for various counties is of little or no significance. The present writer himself confessed in reading the original paper to having some doubts as to the value of the statistical results, but the final verdict on their validity must obviously come from the comparison of the statistical results with the data obtained by the extensive field-mapping of erosion surfaces. It will clearly be many years before erosion surfaces have been mapped over a sufficiently large area of Ireland for such a full comparison to be made, but during the last three years morphological mapping has been carried out by the present writer over more than 1,000 sq. miles of south-eastern Ireland and has included some 700 sq. miles within Counties Wexford and Wicklow, both of which counties were included in the statistical analysis. A full interpretation of the mapped features has yet to be made, and in any case as neither Co. Wexford nor Co. Wicklow has been completely mapped it would be unreasonable to try to compare the results of the mapping so far carried out with the frequency graphs constructed by using all the trigonometrical points in each county. The mapping has, however, demonstrated that flats of apparent erosional origin are widespread in the two counties, and from the work carried out so

far it appears that the flats are grouped at heights which appear to correspond roughly to some, at least, of the peaks in the frequency graphs. The writer thus now feels confident that some of the maxima in the graphs do represent erosion surfaces and is unable to accept Mr. Thornton's conclusion that the maxima occur largely by chance.

With reference to Mr. Thornton's second point that the similarities in the graphs for different counties are again merely chance, the writer can only say that he has traced flats at similar heights over large areas of south-eastern Ireland and has nowhere encountered any evidence suggesting that the platforms have suffered warping. Thus if maxima in the frequency curves for individual counties do represent erosion surfaces it is only reasonable to find maxima occurring at similar heights in adjacent counties.

Mr. Thornton has not been able to examine critically the data on which the statistical analysis was based, and since the appearance of his comments the present writer has compared the distribution of the trigonometrical points used in the analysis with that of the mapped flats. It has been discovered that of the 487 trigonometrical points falling within the mapped areas of Counties Wexford and Wicklow, 353 in fact lie on features mapped as flats. It is thus now clear that the distribution of trigonometrical points is not a random one, and that the Ordnance surveyors unconsciously selected morphological flats as the sites for the majority of trigonometrical points. In view of this the existence of a relationship between erosion surfaces and maxima in the frequency graphs is hardly surprising. It might, of course, be argued that since the recognition of flats in the landscape is subjective, the present writer has, with the earlier statistical work in mind, been tempted to map the areas around trigonometrical points as flats. In fact, however, there is little likelihood of this having happened for the statistical analysis was, for various reasons, based on the trigonometrical points shown on the first edition of the Ordnance Survey's Six Inch Maps, whereas field mapping is being carried out on the current Six Inch Maps which carry a slightly different series of trigonometrical points.

The writer thus finds it difficult to agree with Mr. Thornton that the statistical work was apparently fruitless. When the present programme of morphological mapping in south-eastern Ireland has been completed it is hoped to compare the results of the mapping with the results of the statistical analysis and, if the results of the latter prove valid, to discuss their implications.

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University of Dublin.

Gordon L. Davies.

9th November, 1960.

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REVIEW

LEAD ISOTOPES IN GEOLOGY. By R. D. RUSSELL and R. M. FARQUHAR. Published by Interscience Publishers, Inc., New York (Interscience Publishers, Ltd., London), 1960, pp. 243. Price $9.00.

The best known geological use of the ratios of the isotopes of lead is to the estimation of the ages of uranium and thorium ores. Only the last of the eight chapters of this book is concerned with this. The rest discussess the isotope ratios in lead ores.

Some years ago Holmes and Houtermans showed that on certain assumptions the age of the earth's crust could be estimated from such ratios. Unfortunately the measurements for many ores do not fit into the simple theory and the significance of the results was difficult to judge. Recently the isotope ratios for lead from meteorites has been determined ; by combining these with the terrestrial ratios concordant and apparently reliable results have been obtained and there seems little doubt that the solar system has existed for 4,500 million years.

The authors give a clear explanation of the somewhat confusing theory of these matters, and discuss the origin and possible history of the " anomalous leads ". There has been no easily available and systematic treatment of the matter, and the book is a welcome addition to the literature. The last 116 pages, about half the book, consists of tables of isotope ratios which are believed to give an exhaustive list of all published measurements. They should be sufficient to satisfy the most avid appetite for such information.

E. C. B.

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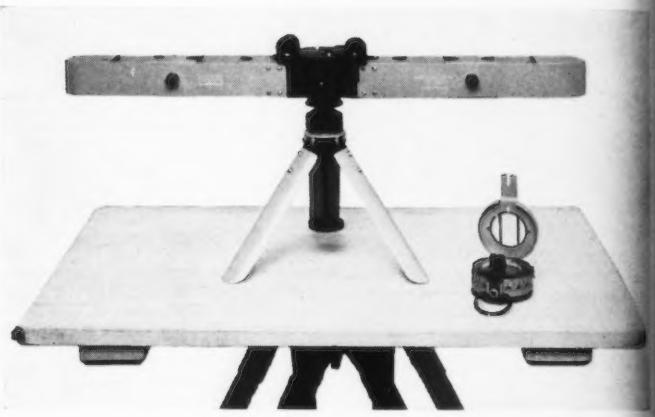

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